ИЗВЕСТИЯ
АКАДЕМИИ НАУК СССР
СЕРИЯ ГЕОЛОГИЧЕСКАЯ

IZVESTIYA AKAD. NAUK SSSR
SERIYA GEOLOGICHESKAYA

1959 No. 2, February

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T-00357. Approved for printing January 26, 1959
Circulation - 3850 copies. Order 3353.
Paper size 70 x 108-1/16. Paper 4. Printing sheets 10.96 + 3 ins. Publ. sheets 11.6

Second printing office of the USSR Academy of Sciences Publishing House
Moscow, Shubinskiy per. 10.

ELBRUS (GEOLOGICAL SKETCH)[1]

by

K.N. Paffengol'ts

This paper gives a detailed petrographic, mineralogic and chemical description of various extrusive rocks of the Elbrus massif, divided here into two sequences. The lower one (up to 2 kilometers thick) has been definitely disturbed and dips generally northeast, at 7° to 10°, with the upper one resting upon it, and making up the upper half of Elbrus; a lava flow is present on its northeastern slope in the Malka valley.

Proof is cited for an Oligocene age of the lower sequence and for an Akchagyl age of the upper. A brief outline of the geologic history of the area is given.

* * * * *

Elbrus is without peer among the lofty peaks of the Caucasus, in its height and majestic beauty. The ancients knew it by many different names; many poetic legends and sagas are connected with it.

The Elbrus area has long attracted explorers of nature; voluminous literature has been dedicated to it; its extrusives have been fairly well described; but up to now there have been no definite data on the age and structure of this volcanic massif.

H. Abich [31] provided the first reliable data on the Elbrus lavas; V.V. Dubyanskiy gave purely petrographic descriptions of the volcano [11, 12]; a detailed geologic survey of the area was made by A.P. Gerasimov [7, 9], S.P. Solov'yev [24, 25, 26] and others. In recent years, comprehensive papers on the area were published by M.V. Muratov [18], M.V. Muratov and M.V. Gzovskiy [19], and Ye.Ye. Milanovskiy [17].

Our own acquaintance with Elbrus dates back to the years between 1913 and 1916, when the author participated in Muratov's geologic party which surveyed the northeastern and eastern foothills of the massif (upper courses of Malka and Irik rivers). After a long interval, the author carried out special geologic studies along the entire Elbrus periphery, during the summers of 1953 to 1957.

The double-headed giant (elevations, 5629 and 5592 m),[2] somewhat to the north (13 km) of the main Caucasian Range looms against it as a comparatively small volcanic structure of the so-called Main Range ("gray") granite, middle Carboniferous in age. Precambrian (?) crystalline schist is exposed along its eastern and southeastern periphery, with Paleozoic deposits to the north (Fig. 1).

The Elbrus dimensions (i.e., area of volcanic rocks) is but 13 x 20 km, with thė cone about two km high. Its morphologic feature is the very small number of valley terrace lava flows, present only in the Malka valley, among the northeastern foothills [7]. There are no traces of "lava flows" which, as mentioned in the literature, are supposed to be located in the Azau, Gara-Bash, Terskol, and Ullu-Kam canyons. The Azau "lava flow" turned out to be ground moraine, but the extrusives of Terskol, Gara-Bash and other areas are definitely disturbed and dip to the northeast, i.e., in a direction opposite to the "flow."

Along the entire periphery of the massif, volcanic rocks are found at comparatively high elevations above the canyon bottoms (at Kyukurtlyu, Azau-Terskol, Syltran-Kyrtyk, Ullu-Kam), locally forming cliffs several hundred meters high. Only in the Biy-Tyk-Tyubyu (on the northwestern slope) and in Terskol canyon does the volcanic sequence reach the bottom, along a major fault (Fig. 1).

STRATIGRAPHY OF THE LAVAS

A.P. Gerasimov, V.V. Dubyanskiy, S.P. Solov'yev and others believed that the Elbrus massif had been built up by three stages of lava flows. The oldest are liparite and its tuff, located at different peripheral points of the volcanic massif; they are followed by andesite-dacite and andesite, followed in turn

[1] El'brus (geologicheskiy ocherk).

[2] The latest maps give 5633 and 5595 m respectively.

by more acid extrusives represented by assorted dacites. According to Dubyanskiy [11], the eastern peak is made up of pantellerite-dacite, and the western peak is made up of hyalopilitic pyroxene-amphibole dacite.

Strangely enough, the disturbances in volcanic rocks at the base of the Elbrus massif are nowhere mentioned in the literature. An explanation is that none of the earlier investigators studied the entire periphery of the massif but rather its individual localities, and all of them cite V. P. Rengarten's data from the Nal'chik area, for the age of the extrusives (Apsheron-post-Pliocene) [23].

We believe from our observations that the Elbrus consists of two extrusive sequences of different ages. The lower sequence, obviously disturbed and greatly eroded, is about two km thick. The upper sequence is thin, resting upon the lower one and forming small flows along the southern slope (the summit belt of V. V. Dubyanskiy) and the large Malka valley flow in the northeastern foothills. The structure of the massif will be considered below.

Basal rocks of the lower volcanic sequence and their contacts with underlying granitoids are well exposed at the foot of the western and south slopes, along a distance of some 25 km from Biy-Tyk-Tyubyu River in the northwest, to the Terskol canyon, in the southeast. Isolated exposures of this sequence have been observed along the eastern slope of Elbrus, on watersheds between Terskol-Irik and Chat-Dzhilky-Augan-Chiran glaciers (Fig. 1).

Major outcrops of similar rocks were observed on the northern slope, between Ullu-Mal'yan-Durku and Mikel'-Chiran glaciers; and farther west, between Kara-Chaul and Ullu-Chiran glaciers, also along the edge of the Irakhit-Syrt plateau. The northernmost exposure of these rocks is located along the northeastern slope of the Tashly-Syrt range, on the left bank of Malka, where they make up the Tuzluk summit (2582 m). Their eastern-most exposures are located on the ridge dividing the Baksan River and its left tributary, Kyrtyk, where the exposed area is 10 sq km. North of there, along the left side of the Kyrtyk gorge, S.P. Solov'yev [25] outlined major outcrops of dellenite (quartz-trachyandesite), hypabyssal varieties of contemporaneous rocks. Rocks of the upper part of the lower Elbrus volcanic sequence crop out in isolated windows among firn and snow fields of the southern slope.

A thick volcanic sequence, well stratified and dipping gently northeast (at angles up to 10^o, Fig. 2), is exposed at the southwestern foot of Elbrus, north and northeast of the Khotyu-Tau pass (3541 m). It was formerly believed to be "flowing" down the Ullu-Kam canyon (i. e., southwest). It consists chiefly of tuffs and tuffaceous breccia, corresponding to liparite-dacite, in composition. Similar rocks are widely developed to the northwest, along the strike, where they make up immense cliffs (up to 1 km high) at the head of the Kyukurtlyu glacier. The interbedded lava is of various hues of gray, superficially rough ("trachyte-like"), and poor in incrustations. On microscopic data and by chemical composition, this lava has been assigned to vitrophyric hypersthene-biotite and hypersthene-biotite-hornblende dacite locally carrying a small amount of resorbed quartz grains. Some varieties exhibit a banded structure of alternating dark and red bands.

Farther on northwest of Kyuturtlyu, layers of dacite lava become more prominent within the clastic volcanic sequence, with the left side of the Biy-Tyk-Tyubyu canyon, in its upper reaches, made up of massive dark-gray vitrophyric hypersthene and hypersthene-biotite-amphibole dacite.

East-southeast from the Khotyu-Tau pass, at the head of Malyy Azau glacier, dense, fine-grained liprite-dacite tuffs are exposed, with hypersthene-biotite, hypersthene-biotite-amphibole and hypersthene-amphibole dacite below them, many of which exhibit a banded texture. Similar rocks are also exposed in the area of the so-called "Leitsinger Lookout," east of the Malyy Azau glacier; they also make up Terskol' peak (3642 m).

Peculiar black-gray lavas with well developed columnar jointing (3 to 8 angles) are exposed stratigraphically and topographically lower, at the base of the lower volcanic sequence, between the Bol'shoy Azau glacier and the lower course of Terskol' canyon, some 400 to 500 m above its bottom. They are overlain by gray, somewhat porous lavas changing upward to agglomeratic and finally to dark, dense varieties, commonly with red oxidized zones. Locally, lavas appear to be banded because of the alternation of red and black colors. On the basis of microscopic study and chemical composition, they are classified as hypersthene-amphibole-biotite dacite. In its northwestern extension, this sequence corresponds to the above-mentioned Khotyu-Tau-Pass rocks.

Lavas rest upon a leveled surface of upper Proterozoic (?) crystalline schist, dipping definitely southwest at an angle of up to 10^o Along the right side of the Terskol gorge, the sequence has been cut off by a major fault and is missing to the east. This fault was formerly believed to be the terminal of a mighty lava flow descending from Elbrus.

This fault has a throw of about one km,

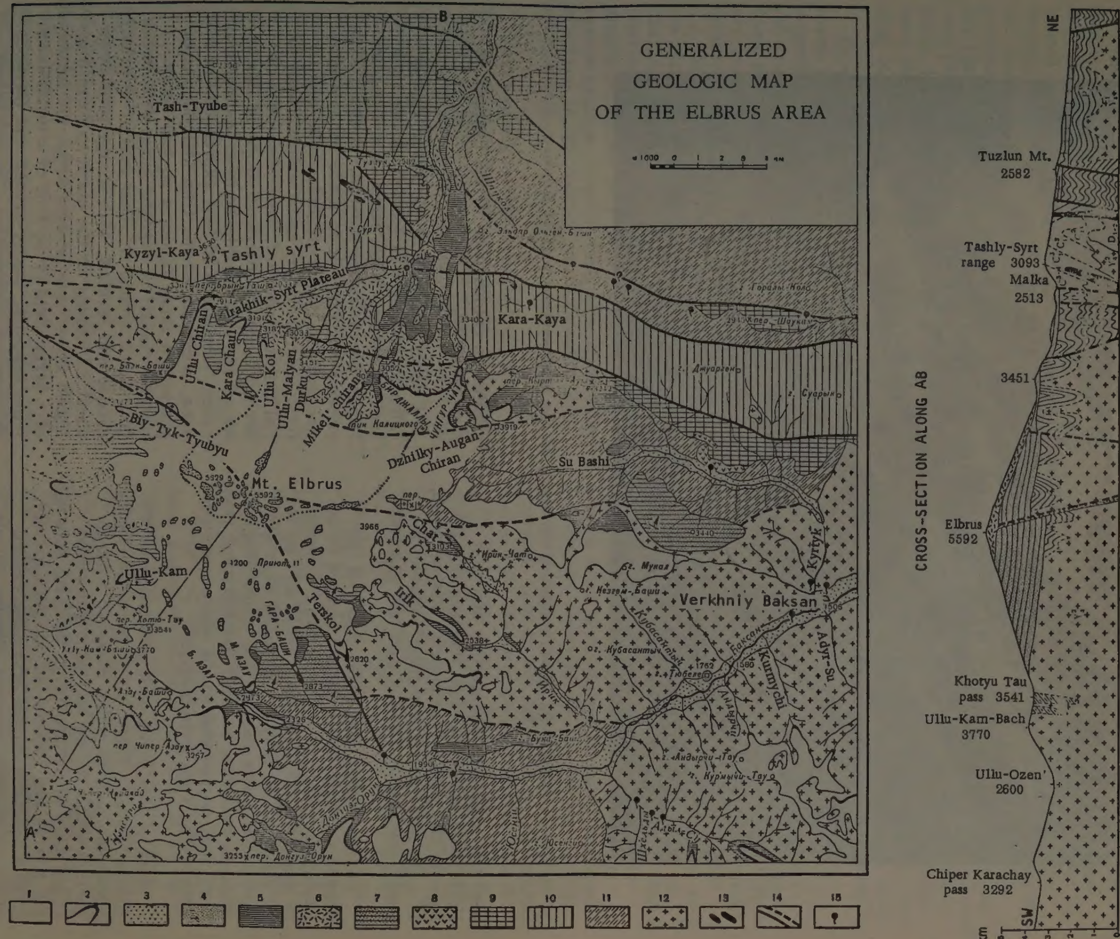

FIGURE 1.

1 -- glaciers, firn and snow fields; 2 -- recent moraines; 3 -- terraces and alluvial fans; 4 -- fluvial-lacustrine deposits; 5 -- ancient moraines; 6 -- lavas of the Malka flow (andesite-dacite and dacite); 7 -- lower (disturbed) Elbrus volcanic sequence: dacite, liparite and their tuffs and tuffaceous breccia; 8 -- dellenite (hypabyssal) of the Kartyk valley; 9 -- upper Liassic arenaceous-argillaceous sequence; 10 -- middle-upper Paleozoic metamorphics; sandstone, limestone, phyllite, shale, quartzite, porphyrite, etc.; 11 -- crystalline schist (upper Proterozoic (?)--lower Paleozoic); 12 -- granodiorite and some migmatite; 13 -- peridotite, etc.; 14 -- normal and reverse faults; 15 -- mineral springs.

FIGURE 2. Gently dipping volcanics at the southwestern foot of Elbrus.

with the west side downthrown and east (ESE) side upthrown. For this reason, the base of the volcanic sequence, farther north, is seen high on the watershed between the upper courses of Terskol and Irik glaciers.

Farther north, the basal rocks of the Elbrus volcanic sequence are located on its eastern periphery, at the western terminal of the Irik-Chat-Kara pass (3642 m), up to an elevation of 3740 m. Exposed here are liparitic (liparite-dacite) tuff, mineralogically close to dellenite; in their stratigraphic position, they correspond to the above-mentioned Khotyu-Tau Pass pyroclastics.

East of that point, and isolated from the Elbrus massif, there are hypersthene-biotite-amphibole dacites which make up a large portion (about 10 sq km) of a watershed ridge (with elevations up to 3400 m) between the Kyrtyk and Syltran-Su. Morphologically, it is a typical remnant of a vast eroded sheet, probably connected with the Elbrus massif. There are no lava flows from its summit and in the direction of the canyons, some investigators to the contrary. The lava is over 400 m thick, with its base 600 to 800 m above the bottom of the Kyrtyk canyon. It is little disturbed and dips northeast.

Immediately north of the dacite butte, hypabyssal liparite-dacite (dellenite) is exposed along the left slope of the Kyrtyk canyon, in two discordant dikes cutting Lower Jurassic sandstone. These rocks are similar in composition to the Pyatigor'ye trachyliparite.

Another remnant of the lower Elbrus volcanic sequence is located on the right bank of Malka, 5 km north of the popular Dzhily-Su carbonate springs, at the Tuzluk summit (2582 m). It is made up of a gray, fine-grained biotite-dellenite tuff. Similar tuffs, correlated by many authors with the Nal'chik liparite tuff, have also been observed along the northern slope of Elbrus where the relationship between ancient and recent extrusives is best expressed. In that area, northeast of the Ullu-Mal'yan-Durku glacier, a large (about 4 sq km) dacite-liparite massif is exposed along with dacite lava flows around it. That massif is greatly eroded and leveled, with bumpy lava flows about it, uniting in the north to form a girdle. These flows have brought about the formation of a large field (1 sq km) of alluvial-lacustrine deposits, partly eroded and terraced. In chemical composition, the extrusives belong to a lime-alkaline branch of liparite-dacite.

All these rocks were first identified by A.P. Gerasimov ([8], p. 43) who has not made a map of the massif. The first geologic map, at a scale of 1:160,000, was made by S.P. Solov'yev [30] who designated the entire massif as dellenite, without giving any description of them.

In 1955, A.L. Lunev first noticed exposures

of a block-type granitoid gneiss, in the western part of the massif. He did not give any description of that rock, but such it has turned out to be. This granitoid gneiss also extends farther east, in a wide band, cutting the massif in two. In the south, the stratigraphic contact between the crystallines and the liparite-dacite gently dipping to the northeast (Fig. 1, cross-section AB) is well defined; in the north, the liparite-dacite has been dropped some 600 m along a fault dipping steeply in that direction.

Farther northwest of that massif, rocks of lower horizons of the lower Elbrus volcanic sequence are exposed on both steep slopes of the Irakhik-Syrt plateau. Here, they appear from under ancient moraine formations, and are represented by banded hypersthene-biotite-amphibole dacite similar to those of the Terskol glacier area. Farther southwest, the cliffs in the left wall at the head of Ullu-Chiran glacier are made up of black-gray vitrophyric hypersthene and hypersthene-amphibole dacite in a fault contact with Hercynian granite. Typical fault breccia has been observed in the contact zone. Granites are mylonitized.

Rocks of higher horizons of the lower Elbrus volcanic sequence crop out in isolated exposures along the southern slope of the massif, where they are represented by hypersthene-biotite-amphibole dacite and liparite-dacite, similar in composition to alkali-earth trachyte (area of "Asylum 11" and "Pastukhov Belvedere"), and exhibiting a banded structure. Still higher stratigraphically and topographically, light gray, finely-porous, locally pumice-like, somewhat alkaline, hypersthene-biotite liparite forms a cornice at the base of the southern slope of the eastern top cone of Elbrus. This is one of the most acid varieties of Elbrus extrusives; it caps the lower ancient volcanic sequence.

Northeast of the subject area, extrusive remnants similar to those at the base of the Elbrus sequence have been found in the basin of Tyzyl and Gundelen rivers (Dzhambash, Kara-Tyube, Yesyuzleruchu – 2253 m, and other peaks). These remnants are links connecting ancient extrusives of the Elbrus massif and the Cheremo-Nal'chik area volcanics. Finally, it is of interest that similar rocks have been found recently in the Western Caucasus, as well, i.e., Tertiary trachyte, trachyliparite and stock-like sub-intrusions of porphyries, in the upper reaches of Pshish River [3].

Some data on the origin of the above-mentioned spotty-banded lavas are pertinent. D.S. Belyankin, who studied them in great detail [4], came to the conclusion that brecciated lava had been formed as a result of fracturing in the movement of a nearly solidified but still hot lava flow, with the vigorous action of gas (air) jets captured by it and oxidizing it. The oxidation process raised the temperature of the corresponding lava segments and lowered their viscosity, thus allowing additional movement and microfracturing, and the assimilation of non-gassed segments by the microbreccia, as a component of the macrobrecciated lava. Belyankin emphasizes the crystallization of silicate lava glass, which accompanied this process, with liberation of silica in the form of cristobalite (tridymite).

This lower volcanic Elbrus sequence, disturbed and eroded, is blanketed by much younger extrusives which make up the Elbrus summit, its northeastern slope, and a major flow along the ancient Malka valley, where it is about 24 km long and 200 m to 1 km wide, in the lower course. The flow is made up of gray to dark-gray hypersthene andesite-dacite and dacite, 87 to 188 m thick (the latter figure for its terminus). The flow which issued from both summit craters of Elbrus, formed the northeastern foot of the massif, and then flowed along the Malka right bank, crossing over to the left bank at the Babugey-Kol mouth. Younger lavas are poorly developed along the south Elbrus slope. The Malka flow lavas commonly exhibit a well-defined columnar parting which becomes locally tabular and unevenly polygonal. Undoubtedly, we deal here not with a single thick flow but rather with a number of superposed flows, rapidly following each other, which accounts for the lack of breaks between them (e.g., a layer of soil or weathering crust). These lavas are marked by their fresh aspect, high density, a near lack of pores and cavities, an abundance of plagioclase incrustations, and a lack of impurities.

Rocks of the Elbrus summit cones have been described in most detail by V.V. Dubyanskiy [12] who personally studied them during 1907 to 1913. The western summit rocks have been described also by F. Shafarzik (from a collection of M. Dechy, [37]) and by L. Ammon (from G. Merzbacher collection, [32]); the eastern summit rocks -- by S.P. Solov'yev [27] from his personal collection.

In the area of the western summit cone, V.V. Dubyanskiy recognizes lavas of the summit crest and its beak-like protrusion, also scoriaceous lava and pumice at its foot and the lava of near-summit rocky crests. These lavas are vitrophyric hypersthene-biotite-amphibole dacite, andesite-dacite and somewhat alkaline liparite-dacite.

In the eastern summit cone area, V.V. Dubyanskiy [12] discovered several varieties

of hypersthene dacite, their tuff and tuffaceous breccia, also ash (clinker sand). In summit parts of the crater funnel, dark-gray lavas of various hues are exposed, commonly porous and rich in feldspar incrustations (up to 8 mm). This ash belongs to the alkaline branch of liparite and liparite-dacite. The pyroclastic formations are the youngest present in the Elbrus massif.

The same age should be assigned to a lava flow, 50 to 70 m thick, on the northern slope of the Tashly-Syrt range, at the Khudes headwaters. Their source must have been the flat, shield-like Tash-Tyube cone (2479 m). According to N.I. Tsibovskiy, they are represented by biotite-hypersthene and hypersthene andesite and andesite-dacite.

CHEMISTRY OF THE LAVAS

An idea of the chemical composition of the Elbrus lavas may be gained from the fairly numerous isolated analyses given by different authors, mainly by Dubyanskiy [12]. We have systematized these analyses and supplemented them with two new ones (Nos. 10 and 26), carried out on our own specimens (1957). All analyses have been converted into A.N. Zavaritskiy's numerical designations.

A petrographic study of both extrusive sequences has shown that they were formed as a result of magmatic cycles of different ages similar in composition. The acidity in both cycles increases toward the top. The older cycle opened with dacite and terminated with liparite, but the following cycle opened with an outflow of andesite and andesite-calcite (Malka flow) and ended with dacite and liparite-dacite (near-summit flow).

To be sure, there are certain differences in the chemical and mineralogic composition and the microstructure of lavas, but on the whole they belong to the same broadly defined dacite type, and are marked by high acidity. Alkaline varieties predominate (see diagram Fig. 3) but never go beyond the lime-alkaline rock association [14]. There are both typical dacite and varieties transitional to andesite as well as to trachyte, liparite, and rhyolite. Most rocks usually are somewhat supersaturated with silica, which appears to be due to its presence in the dark-colored components. All of the rocks appear to be close in alkali content and in Na/K ratio (all their vectors are nearly parallel, as witness the left side of diagram Fig. 3). All rocks are more basic than the average liparite and, in their bulk, much more acid than trachyte and alkaline trachyte.

The average composition of upper lavas (analysis 3) corresponds to dacite, with a certain deviation toward andesite-dacite and a somewhat higher silica content. The average composition of the upper sequence (analysis 34) is similar to that of low-alkaline liparite-dacite.

An outstanding feature of the Elbrus extrusives is their abundance in hypersthene and the predominance of Na_2O over K_2O, even in those solitary instances where the weight percentage of Na_2O is somewhat lower than for K_2O. This makes it possible to regard the subject area as a separate petrographic province of the Caucasus, characterized by a dacite-type mother magma. If this is true, the diversity of the Elbrus volcanics has been brought about by complex syntetic liquefaction and differentiation of magma in its hearth, during a period of quiescence prior to eruption, and partly by the processes of fractional crystallization, which took place in isolated sectors of a magma already split up, at a temperature close to that of solidification, or possibly even during the crystallization (as witness the splitting up in two layers, of lava in the upper belt of the western summit cone, according to V.V. Dubyanskiy).

Judging from their chemical analyses, the Elbrus volcanics belong to derivatives of magmas relatively rich -- for a given silica content -- in lime, magnesium and iron, but somewhat impoverished in alkali.

In Zavaritskiy's classification [14], these rocks correspond to the lime-alkali association which comprises rocks ranging from the island of Martinique and ending with the Yellowstone lavas.

STRUCTURE OF THE MASSIF

The Elbrus massif is made up of two extrusive sequences sharply differing in their morphology. The lower volcanic sequence, which accounts for the bulk of the massif, is obviously dislocated and strongly eroded. The younger lavas rest upon it, mantle-like, and make up the flow down the northern slope and the foothills, spreading farther north for 12 km more along the Malka valley.

Both extrusive sequences were formerly regarded as quasi-contemporaneous, having issued from a common source along the entire periphery of the massif. The small extent of these older flows was explained by a high viscosity of lava, which, in addition, was thought to be responsible for the formation of lava falls.[1] As a matter of fact, the

[1] Ice falls were also connected with them, although ice falls have also been observed on erosional escarpments (Irik, Terskol, etc; see Fig. 4).

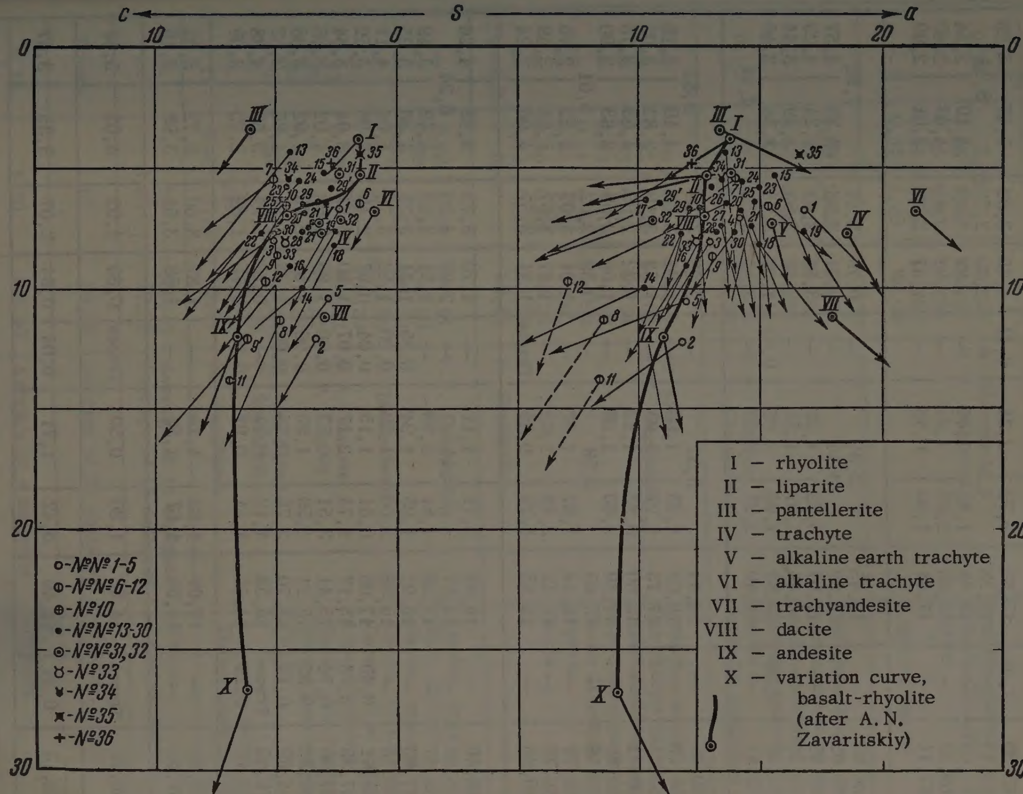

FIGURE 3. Chemical diagram of the Elbrus area extrusives.

1-5 -- western summit lavas; 6-12 -- eastern summit lavas; 10 -- lava of northern foot of massif; 13-30 -- rocks of lower volcanic sequence; 31-32 -- dellenites (hypabyssal); 33 -- average composition of upper lavas; 34 -- average composition of lower lavas; 35 -- average composition of the Pyatigor'ye trachyliparite; 36 -- average composition of the Ullukam granite.

Composite table of analyses of the Elbrus area extrusives

Analysis no.	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O
1	63,80	1,04	15,62	0,72	2,31	0,11	1,15	3,32	5,47	3,26
2	65,14	—	17,28	4,54		—	2,13	2,72	6,14	
3	65,61	—	18,48	1,81	1,44	—	1,89	4,06	4,19	2,54
4	66,23	—	17,00	2,63	1,31	—	1,82	3,85	4,63	2,80
5	67,11	—	17,51	1,96	0,48	—	1,37	2,29	3,61	2,69
6	66,89	—	13,45	3,94		—	0,79	2,41	7,92	
7	66,02	—	18,59	3,14		—	0,71	3,88	3,23	4,57
8	68,92	—	17,07	0,72	1,31	—	3,10	4,01	3,24	1,29
9	65,34	—	18,46	1,91	1,55	—	2,16	3,95	3,79	3,44
10	66,50	0,83	15,89	1,93	2,02	0,03	1,69	3,74	3,74	3,40
11	62,42	—	18,40	2,23	3,46	—	3,19	5,35	4,47	
12	69,37	—	14,44	5,32		—	2,26	4,38	3,82	
13	66,06	—	16,61	1,33	1,08	—	1,32	3,66	4,10	3,18
14	68,31	—	17,73	1,27	1,06	—	1,53	3,19	2,82	3,05
15	69,41	—	15,60	0,83	1,82	—	1,32	2,95	4,63	3,75
16	66,85	—	17,10	2,80	1,28	—	1,96	3,58	3,63	2,90
17	73,58	—	13,65	3,58		—	0,39	2,02	5,61	
18	66,79	—	15,18	1,87	1,07	—	1,91	3,97	5,22	2,76
19	66,76	—	17,12	2,78	1,05	—	1,67	3,67	5,64	3,49
20	65,85	—	15,85	4,24	0,35	Traces	1,21	3,57	4,96	2,24
21	66,86	—	16,04	2,71	1,11	—	1,70	3,73	3,89	4,26
22	65,84	—	17,67	3,44		—	2,02	4,39	6,34	
23	65,67	—	17,58	1,81	0,98	—	1,91	3,69	4,49	3,49
24	67,80	—	16,92	1,05	1,94	0,35	1,31	3,25	4,36	3,35
25	65,75	—	18,38	2,60	1,30	0,20	1,52	3,70	4,04	4,11
26	69,42	0,45	17,05	0,81	1,15	0,04	0,97	2,40	3,98	3,52
27	67,35	0,76	15,17	1,74	2,17	0,07	1,80	3,54	3,94	3,34
28	67,06	0,74	15,41	3,51	0,57	0,06	1,85	3,69	4,02	3,05
29	68,32	0,52	15,21	1,59	1,95	0,06	1,51	2,99	2,92	3,96
30	64,69	0,73	18,70	3,76	0,78	0,09	1,23	3,91	4,66	2,52
31	71,46	Traces	16,58	1,13	0,91	—	0,22	1,91	3,60	4,38
32	72,12	0,28	14,46	1,41	0,97	0,03	0,75	1,82	2,95	3,09
33	65,81	—	17,07	1,82	1,79	—	1,82	3,60	3,72	3,02
34	67,52	—	16,35	1,67	1,33	—	1,58	3,54	3,98	3,16
35	69,73	Traces	14,56	1,80	0,50	Traces	0,84	1,92	4,02	5,59
36	72,15	0,31	14,49	0,42	1,71	0,04	0,64	2,07	3,35	3,47
37	72,90	0,48	14,18	1,65	0,31	0,13	0,40	1,13	3,54	3,94
38	68,63	0,35	10,30	5,60	2,61	0,21	0,37	1,07	6,14	4,17
39	65,68	0,57	16,25	2,38	1,90	0,06	1,41	3,46	3,97	2,67
40	63,91	0,59	15,88	3,22	2,23	0,01	1,14	2,81	3,08	5,80
41	59,59	0,77	17,31	3,33	3,13	0,18	2,75	5,80	3,58	2,04

Note: Comma represents decimal point.

Composite table of analyses of the Elbrus area extrusives, continued

P_2O_5	H_2O^-	H_2O^+	Sum	Analyst	Rock
0,19	3,59		100,58	A. Shvager	Andesite-dacite
—	0,86	1,19	100,00	V. V. Dubyanskiy	Pumice (dacite)
—	0,15	0,37	100,54	"	Dacite
—	0,11	1,22	101,60	"	"
—	0,35	1,69	99,06	"	Liparitic dacite
—	1,69	2,91	100,00	"	Tuff (pantellerite-dacite)
—	0,41	2,10	102,65	"	Scoriaceous sand (liparite-dacite)
—	0,14	0,86	101,00	"	Liparite-dacite
—	0,13	0,41	101,14	"	Andesite-dacite
0,18	0,26	0,30	100,24	V. A. Yusova	Dacite
—	0,48		100,00	V. V. Dubyanskiy	Andesite (inclusion in No. 9)
—	0,60		100,19	H. Abich	Liparite-dacite
—	0,13	0,56	101,03	V. V. Dubyanskiy	Pantellerite-dacite
—	0,03	0,55	99,54	"	Liparite-dacite (vein in No. 13)
—	0,49	0,67	101,47	"	Pantellerite-dacite
—	0,21	0,68	100,99	"	Dacite (banded)
—	0,04	1,13	100,00	"	Liparite
—	0,16	0,32	99,25	"	Dacite
—	0,13	0,35	102,66	"	"
CO_2—0,39	0,02		99,32	F. Shafarzik	"
—	0,03	0,08	100,41	V. V. Dubyanskiy	Dacite, gray
—	0,09	0,21	100,00	"	Tuff, dacitic
—	0,17	1,62	101,39	"	Dacite, mottled
—	0,33		100,66	A. Dannenberg	Dacite, (black variety)
—	1,20		102,20	"	Dacite, (red variety)
0,11	0,10	0,32	100,32	V. A. Yusova	Liparite-dacite
0,32	—	0,11	100,31	K. P. Sokova	Dacite (black variety)
0,32	0,08	0,07	100,34	"	Dacite (red variety)
1,10	0,40	0,11	99,64	M. Vasil'yeva	Dacite
—	0,22		101,29	F. Shafarzik	"
—	0,32		100,52	"	Liparite-dacite (alkaline branch)
0,11	0,40	1,12	99,51	M. Vasil'yeva	Dacite-liparite (dellenite)
—	1,75		100,40	Average of 11 analyses	Dacite
—	0,55		99,68	Average of 18 analyses	Liparite-dacite
Traces	1,02		99,98	Average of 35 analyses	Trachyliparite
0,15	1,20		100,00	Average of 4 analyses	Granite
0,01	1,33		100,00	Computed	Liparite
0,02	0,53		100,00	"	Pantellerite
0,15	1,50		100,00	"	Dacite
0,05	1,28		100,00	"	Trachyte, alkaline earth
0,26	1,26		100,00	"	Andesite

Note: Comma represents decimal point.

Conversion of analyses by the Zavaritskiy method

Analysis no.	Numerical Characteristics after A. N. Zavaritskiy										
	a	c	b	s	a'	f'	m'	c'	n	φ'	t'
1	16,8	2,2	6,7	74,3	—	42,2	28,9	28,9	72,1	8,2	1,2
2	11,8	3,2	12,1	72,9	37,6	33,1	29,3	—	76,1	17,7	—
3	12,9	4,9	8,0	74,2	24,0	35,9	40,1	—	71,5	19,0	—
4	14,1	4,3	7,0	74,6	—	49,6	44,4	6,0	71,1	31,7	—
5	11,7	2,7	10,4	75,2	57,1	20,1	22,8	—	66,6	15,6	—
6	15,3	1,4	6,5	76,8	—	54,2	21,3	24,5	71,1	36,1	—
7	14,0	4,9	5,4	75,7	27,8	50,6	21,6	—	51,4	33,0	—
8	8,6	4,7	11,3	75,4	34,6	15,0	50,4	—	78,7	4,6	—
9	13,0	4,7	8,6	73,7	22,0	35,4	42,6	—	62,5	18,8	—
10	12,6	4,4	6,6	76,1	—	54,1	43,9	2,0	60,8	25,0	1,0
11	8,4	6,6	13,8	71,2	22,9	38,0	39,1	—	65,6	13,9	—
12	7,1	5,3	9,7	77,9	12,5	48,7	38,8	—	70,0	29,1	—
13	13,5	4,2	4,3	78,0	—	47,0	50,0	—	66,0	25,0	—
14	10,3	3,8	9,9	76,0	54,0	20,2	25,8	—	58,4	10,8	—
15	15,6	2,7	5,3	76,4	—	44,3	40,5	15,2	65,5	12,6	—
16	12,0	4,3	9,0	74,7	23,1	40,0	36,9	—	65,2	26,1	—
17	10,3	2,4	6,3	81,0	40,0	50,0	10,0	—	70,5	31,5	—
18	15,1	2,4	8,1	74,4	—	32,2	38,8	29,0	74,3	20,0	—
19	16,8	2,3	7,6	73,3	—	42,6	34,8	22,6	71,1	29,6	—
20	14,1	3,6	6,8	75,5	—	57,5	31,3	11,2	77,7	52,5	0,7
21	14,6	3,2	7,3	74,9	—	45,0	38,5	16,5	57,8	31,2	—
22	11,8	5,4	7,6	75,2	16,2	38,7	45,1	—	62,7	21,6	—
23	15,0	4,4	5,7	74,9	—	42,4	55,3	2,3	66,1	25,9	—
24	14,2	3,9	5,5	76,4	7,4	53,1	39,5	—	66,6	14,8	—
25	14,8	4,5	6,3	74,4	10,7	48,4	40,9	—	59,6	25,7	—
26	13,6	2,8	6,4	77,2	47,3	26,8	25,9	—	63,7	10,3	0,5
27	13,3	3,4	7,4	75,9	—	48,7	40,5	10,8	64,6	19,9	0,8
28	13,3	3,7	7,5	75,5	10,0	48,2	41,8	—	66,3	40,0	0,8
29	12,1	3,7	6,6	77,6	12,0	49,0	39,0	—	52,8	20,9	0,5
30	14,0	4,8	7,6	73,6	19,7	53,6	26,7	—	73,5	42,8	—
31	13,9	2,2	5,2	78,7	61,0	33,0	6,0	—	55,2	17,7	—
32	10,7	2,1	7,1	80,1	49,3	32,7	18,0	—	59,2	16,8	0,2
33	12,5	4,4	8,0	75,1	20,5	40,5	39,0	—	65,2	19,0	—
34	13,5	4,3	5,4	76,8	—	48,1	49,3	2,6	65,6	25,3	—
35	16,6	1,3	4,4	77,7	—	44,0	31,8	24,2	52,4	33,3	—
36	12,2	2,5	4,8	80,5	36,2	41,6	22,2	—	59,1	8,3	0,3
37	12,9	1,3	5,3	80,5	52,5	35,0	12,15	—	58,2	27,5	0,5
38	13,4	5,8	3,4	77,4	—	45,1	17,7	37,2	55,4	0,0	0,45
39	12,8	4,3	7,0	75,9	9,8	55,9	34,3	—	68,9	29,4	0,6
40	15,5	3,0	7,3	74,2	—	67,6	26,7	5,7	44,6	38,1	0,8
41	11,1	6,4	12,0	70,5	—	51,8	40,6	7,6	75,3	24,7	1,0

Note: Comma represents decimal point.

Conversion of analyses by the Zavaritskiy method, continued

Q	a/c	Locality	Reference
12,8	7,6	Western summit	H. Merzbacher ([36], v 2, p 783)
19,0	3,7	" "	V. V. Dubyanskiy [12]
17,7	2,6	" "	" [11]
16,7	3,2	Same, protuberance	" [12]
24,3	4,3	Same, slope	" [12]
21,6	11,0	Eastern summit	" [12]
22,5	2,8	" "	" "
28,9	1,8	" "	" "
16,7	2,7	Eastern summit, Achker'yakol' flow	" "
23,2	2,8	Northern foot	K. N. Paffengol'ts (1957)
19,0	1,3	Eastern slope, Achker'yakol' flow	V. V. Dubyanskiy [12]
36,3	1,3	Eastern slope	H. Abich ([31], p. 49)
24,8	3,2	Eastern summit, southern slope	V. V. Dubyanskiy [12]
27,6	2,7	" " " "	" "
18,9	5,8	Southern slope, Pastukhov Belvedere	" "
21,1	2,8	" " near-summit belt	" "
39,0	4,3	" " "Asylum 11"	" "
16,2	6,3	Same, upper cliffs	" "
10,7	7,3	Tereskol peak	" "
19,2	3,9	Crest at the Bol'shoy Azau glacier	M. Dechy ([34], v. 3, p. 241)
17,4	4,6	Khotyu-Tau Pass area	V. V. Dubyanskiy [12]
21,4	2,2	Azau glacier	" "
15,4	3,4	Lookout	" "
20,5	3,6	Northwestern foot	A. Dannenberg [33]
14,7	3,3	" "	" "
24,4	4,8	Same, massif at Ullu-Mal'yan-Durku glacier	K. N. Paffengol'ts (1957)
27,2	3,9	Baskan valley, boulders in Tegenkeli area	D. S. Belyankin [4]
21,7	3,6	" " " " " "	" "
27,3	3,2	Head of the Irik glacier, right slope	S. P. Solov'yev [26]
14,6	2,9	Northwest of Urusbiyevo (Upper Baksan)	M. Dechy ([34], v. 3, p. 231)
27,4	6,3	Kartyk gorge	" " " " p. 251
36,7	5,1	" "	S. P. Solov'yev [25]
20,8	2,8	Average composition of upper lavas	K. N. Paffengol'ts
22,3	3,1	" " " lower "	"
20,9	12,7	Pyatigor'ye	A. P. Gerasimov [10]
34,1	4,9	Upper course of Ullu-Kam River (right Kuban' summit)	V. P. Belikov (1948)
33,9	9,9	Average types, after R. Deli	A. N. Zavaritskiy (1947)
22,2	2,3	" " " " "	" "
21,9	3,0	" " " " "	" "
14,4	5,1	" " " " "	" "
12,4	1,75	" " " " "	" "

Note: Comma represents decimal point.

lava falls are typical erosional escarpments and shelves in the disturbed lower volcanic sequence (shelves above the west side glaciers, Biy-Tyk-Tyubyu, Kyukurtlyu, and Ullu-Kam; and in the Azau glacier area, on the southern slope).

Our detailed observations along the entire periphery of the massif have proved that the well-stratified lower volcanic sequence definitely dips SW, at angles up to 10°. Structurally, the massif consists of two tectonic blocks (Fig. 1), roughly of the all-Caucasian trend, divided by a major fault, with the southwestern block downthrown. The northeastern block has been complicated by a number of transverse faults of smaller magnitude. They intersect the main fault and are responsible for a horst and two grabens (Fig. 1 and cross-section AB).

The lower volcanic sequence (the Elbrus massif proper) rests upon an eroded, leveled-off surface of "middle Range granite," i.e., on a rigid crystalline basement. Therefore, the above-mentioned dip of beds should have been caused by faulting. The largest fault has been identified at the northwestern foot of Elbrus, south of the Balk-Bashi pass (3682 m), at the head of the Ullu-Chiran glacier. We have traced that fault over 20 km farther west, along the right side of the Ullu-Khurzuk canyon. Its throw is not less than one km. Because of the definite northeasterly dip, the stratigraphic contact of volcanics with the underlying granite rises south-southeast of the Biy-Tyk-Tyubyu mouth, across the Kyukurtlyu and Ullu-Kam canyons and finally the Khotyu-Tau Pass (3541 m) area, at the southwestern foot of Elbrus.

The contact gradually descends, somewhat north of the Khotyu-Tau pass and farther east-southeast, by virtue of the northeasterly dip of the extrusives. After having passed along the left slope of the upper Baksan valley, it reaches the lower course of the Terskol canyon, where the volcanic sequence is cut off by a northwesterly trending fault which we have first identified. This fault runs along the right bank of that canyon, where the northeastern block also has been raised along a fault with a throw of about 1.5 km. Another major latitudinal fault trends north-northeast in the Kyrtyk range, where it separates granite and crystalline schist, with the fault plane dipping northeast at 55° to 60°, in the Chat glacier area. This fault is undoubtedly older than the lower volcanics, because it is covered by them in the eastern part of the Kyrtyk range (remnant north of the Syltran-Gel' lake). The westerly projection of this fault, however, passes through the Elbrus crater and the so-called Achker'-yakol side crater. It is, therefore, reasonable to assume that the fault has been rejuvenated here and has participated in the uplifting of the Malka flow and the summit

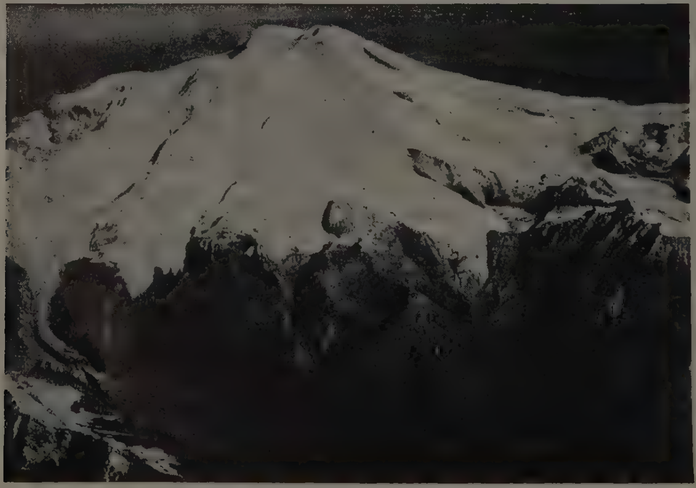

FIGURE 4. Lower Elbrus volcanics and underlying granitoids. Southeastern slope.

belt lavas. It appears that the eastern summit is a junction of the Ullu-Khurzuk and Tereskol faults (Fig. 1). Thus, the Elbrus volcano is located at the intersection of two normal faults of different ages, the older (?) of which has been subsequently rejuvenated.

The Elbrus summits are typical volcanic cones separated by a saddle (at 5300 m) which is a merger of their slopes. The volcano itself apparently was not formed all at once but in several stages. Its two summits, most probably, also are not contemporaneous, as witness the character of their lavas and the morphology of the two cones.

AGE OF LAVAS

The age of the two extrusive sequences of Elbrus may be only indirectly determined. All investigators were unanimous in correlating the more acid volcanics with the Nal'chik region liparite tuff and lavas, Apsheronian in age, according to V. P. Rengarten [23]. The onset of Elbrus volcanism was assigned, then, to the Akchagylian-Apsheronian, and the end -- to the Würm.

In our special paper [21], we considered the age relationship between the Nal'chik extrusives and hypabyssal laccolith rocks of Pyatigor'ye as well as the Eldzhurtin granodiorite, and we proved an Oligocene age for the first and a Lower Miocene for the second. G.D. Afanas'yev [1], who held to a different opinion (1950-1954), has come to share our idea of the Tertiary age of the Eldzhurtin granite, dellenite and trachyliparite.

Papers of G.D. Afanas'yev [2], Yu.P. Masurenkov [16] and G.M. Zaridze in cooperation with Ye. Ye. Milanovskiy [15], which have a direct bearing on our problem, were published in the Izvestiya of the Academy of Sciences, U. S. S. R., Geological Series, No. 6, 1957. It must be emphasized that Afanas'yev's conclusions, while in accord with our own, are in opposition to the rest of the above-named authors.

Of great interest is the frequent occurrence of anorthoclase granite xenoliths (petrographically similar to the Eldzhurtin porphyritic granite) among trachyliparite of Dzhutsa and Zolotoy Kurgan. The absolute age of the latter, as determined by Afanas'yev, is 25 million years, with 35 million years for the granite xenoliths in them (1956 data). In geologic chronology, these figures correspond to the lower Miocene and upper Oligocene.

Yu. P. Masurenkov [16] and others, working southeast of Tyrna-Auz, between Baksan and Chegem rivers, along the lower course of Dzhungu-Su (left tributary of the Chegem), discovered an intrusion of anorthoclase granodiorite porphyry. That rock, according to Afanas'yev ([2], p. 39), is petrographically identical with vein granodiorite porphyry associated with the Eldzhurtin porphyritic granite. It is noteworthy that the intrusion is associated with the structural zone of the Eldzhurtin porphyritic granite. Unfortunately, Afanas'yev does not mention that this intrusion (3 to 4 sq km) cuts, according to Yu. P. Masurenkov ([16], pp. 63-64, Fig. 6) the Baksan-Chegem interfluvial volcanics which he assigns to the lower Pleistocene while he believes the intrusion to be Mindel-Riss in age, which is of course impossible.

It is of interest that farther south-southwest, similar rocks make up a sill [21] in the same volcanic sequence. G. M. Zaridze and Ye. Ye. Milanovskiy believe this sill to be an ancient moraine ([15], p. 105). Their division of the volcanic sequence into two formations of different age is artificial.

Zaridze and Milanovskiy, while denying that the Eldzhurtin granodiorite cuts the Baksan-Chegem watershed volcanics, say nothing of the highly alkaline anorthoclase granodiorite porphyry which cuts this sequence a short distance to the east, in the lower course of Dzhungu-Su.

The two authors believe the south Kyugen-Kaya summit is an eroded lower Tertiary volcanic cone, with lava dipping periclinally along its slope, with remnants of a lava flow present along the Kestanta canyon, on its right bank, 12 km north of the mouth of Khasta River, about 400 to 500 m above the bottom. We have not come across any periclinal dips in lava; the "remnant of a lava flow" we have found to be a Toarcian volcanic facies.

Yu. P. Masurenkov [16] attempts to determine the age of the volcanids by geomorphologic means. However, his basic premise of their relationship with the fourth peneplane is erroneous. According to our observations, this surface (terrace) adjoins the volcanic sequence rather than being covered by it, which invalidates all of Masurenkov's conclusions. The intense disturbance of upper Tertiary pebble beds, which he mentions ([16], p. 60), is a result of landslides. His geologic map of a sector of the Chegem-Baksak watershed is too generalized and incorrect in many instances. The radical divergence in the dips of volcanics, as determined by Masurenkov on one side, and by Milanovskiy and Zaridze on the other [15] is of interest, inasmuch as it obviously reflects preconceived notions on the structure of the Kyuren-Kaya — Kum-Tyube massif.

Yu. P. Masurenkov believes the complete correlation of the Lower Chegem and Upper Chegem areas to be premature, because only the liparite rocks are similar in the lower parts of the sequences, in both areas. Curiously enough, Zaridze and Milanovskiy [15] came to the opposite conclusion -- that the upper part of the Baksan-Chegem watershed volcanics are similar to the lower part of the Nal'chik sequence. As we see it, there is a general northeasterly change in volcanic facies, away from the eruptive centers (?), and toward a junction with marine facies. Masurenkov's hypothesis of an eruption mechanism with an enormous number (6760!) of volcanic centers is not confirmed by the distribution study of volcanic facies of the area, and is purely imaginary.

Until very recently, all investigators were unanimous in correlating the Chegem-Nal'chik extrusives with hypabyssal trachyliparite of the Pyatigor'ye laccoliths. Now, Masurenkov ([16], p. 67) regards this correlation, based on an alleged petrographic similarity, to be a misunderstanding. He emphasizes that the Pyatigor'ye intrusions fall into a comparatively narrow age interval, Karagan-Akchagyl, but the Chegem-Nal'chik extrusives are lower Tertiary.

That author, however, did not take into consideration the fact that the Lysogorsk Plateau trachyliparite does not cut the Karagan deposits; also, according to N. P. Luparev, conglomerate with trachyliparite pebbles turned out to be Meotian rather than Akchagylian, which is in accordance with Afanas'yev's data.

Our detailed study along the right bank of Baksan, in the vicinity of Zayukovo village, in the summer of 1957, has demonstrated that the volcanic sequence unquestionably dips under the Upper (?) Maikop deposits.[1] Accordingly, the age of the lower Elbrus volcanics is most probably Oligocene.

The following field data have bearing on the age determination of the upper Elbrus extrusives. The Elbrus lava flows rest everywhere directly upon ancient stratified rocks. All investigators emphasize that these lavas are overlain by glacial deposits, and never the other way around [7, 9]. It follows that the younger Elbrus volcanism antedated the Caucasian glaciation. The Malka flow lavas fill up the relief of the ancient topography in both Paleozoic and Jurassic rocks, as well as in the lower extrusives. According to Nikolayev [20], the Central Caucasus relief originated in lower or middle Tertiary. An analysis of our cross-sections along the Malka, within the lava-flow area (Fig. 5) reveals the depth of the river bed, 234 to 277 m, below the Dzhyly-Su spring, which exceeds the relative height of the Upper Apsheronian (Gunz) terrace. In the northeastern part of the northern Caucasus, volcanic material has been observed throughout the entire Akchagyl section, being especially abundant in the southern part of the basin [6]. Contemporaneous magmatism coincided with great pre-Akchagylian orogenic movements which brought about the Akchagyl transgression. On the basis of these data, the age of the upper Elbrus extrusives (Malka flow) is most probably Akchagylian.

Different opinions were voiced by M. V. Muratov in cooperation with M. V. Gzovskiy [19] and subsequently by Ye. Ye. Milanovskiy [15]. A structural map by the latter author ([19], Fig. 3, p. 77) shows the Malka flow lavas at the base of the section, and assigned to the Riss-Wurm. The Irakhik-Syrt extrusives are also assigned to the Riss-Wurm, although they overlie the Malka lavas, and the near-summit extrusives are supposed to be post-Wurm. This stratigraphic sequence is obviously incorrect, because there are no lava flows along the Irik, Azau, and Terskol valleys, as has been pointed out above; at the same time, the Malka flow lavas override those of the Irakhik-Syrt plateau, i.e., their position is reversed. In the age determination of lava flows, from their relations to the adjacent and overlying moraines, Muratov and Gzovskiy based their opinion on the erroneous premise that all terraces rise progressively upstream. Accordingly, they assign to the Riss the erosional surface which is marked by ledges with moraines, at 250 to 300 m above the bottom, in the Malka upper course. We regard them as Gunz. This topic will be discussed in more detail, below.

Ye. Ye. Milanovskiy [15] differentiates the peripheral Elbrus andesite-dacite lavas into middle upper Pleistocene and Holocene, dividing the latter into two generations. Our criticism, above, is fully applicable to these conclusions, and our reasons for assigning an Akchagylian age to the younger Elbrus lavas remain valid.

ANCIENT GLACIATION

At the present time, Elbrus is an isolated center of extensive glaciation (about 140 sq km). The central firn field of the massif, 3200 to 3900 m high, sends out 16 comparatively large glaciers and a number of smaller

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[1]K. N. Paffengol'ts does not take into consideration the local volcanic vents through which the extrusives reached the surface. They are especially noticeable in the Zayukovo area. Russian Editor.

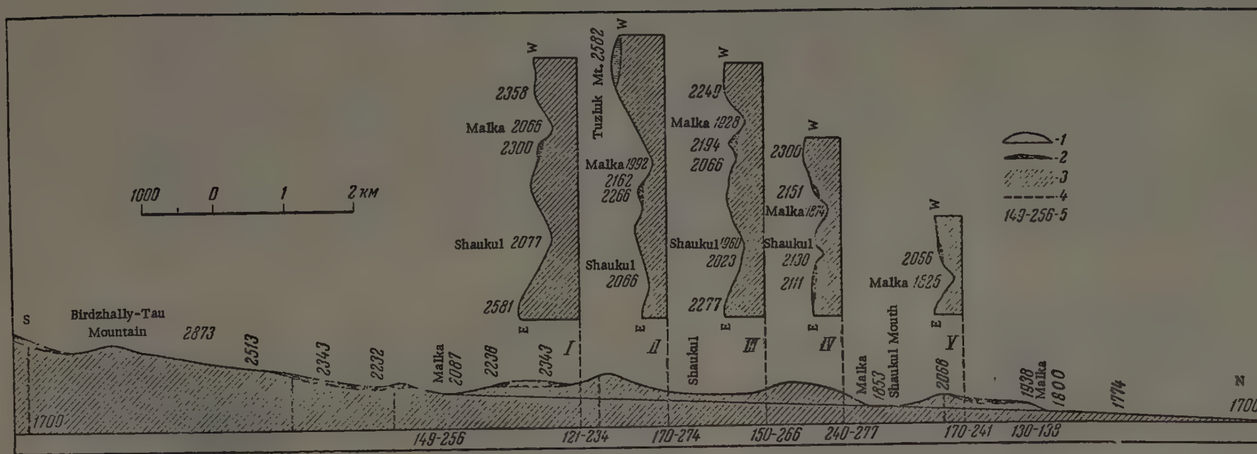

FIGURE 5. Longitudinal and transverse cross-sections of the upper Malka valley, illustrating the depth of river bed cut after the lava eruption.

1 -- ancient extrusives; 2 -- Malka flow lavas; 3 -- underlying rocks; 4 -- normal faults 5 -- depth of cut from the flow base (149 m) and from top (256 m).

ones (Fig. 1), which feed the headwaters of the Kuban, Baksan and Malka rivers. The glaciers, all comparatively short, form small appendices to that vast firn expanse. Elevations of the terminal glaciers are as of the beginning of this century, from the well-known K. I. Podozerskiy's resumé of Caucasian glaciers [22]. All glaciers are now in a recessive stage. This topic is dealt with in special papers of S. P. Solov'yev [28], N. A. Bush [5], G. Distel [35], and others. We are interested here only in the ancient glaciation, insofar as it has a bearing on the age of the younger Elbrus extrusives.

Ancient moraines are widely developed along the entire periphery of Elbrus; ancient glaciers descended long distances along the Malka, Kuban, and Baksan headwater canyons.

Ancient glaciers covered the entire northern Elbrus slope and reached the southern slope of the latitudinal Tashly-Syrt Range. This ice field fed the main glacier which descended farther north, down the Malka valley (down to elevation between 1920 and 2050 m), with branches to the west, toward the headwaters of Chemart-Kol and Ullu-Khurzuk, and over the Bryn-Tash (at 3074 m) and Balk-Bashi (at 3682 m) passes, respectively.

No less than three ancient lateral moraines are present along the higher left slope of the Ullu-Chiran glacier valley, with the upper moraine running south of the Bryn-Tash pass, toward the upper Chemart-Kol canyon (a tributary of the Kuban). Farther down the canyon, at the confluence of two branches of the river, a well-defined terrace between 180 and 250 m above the bottom is made up almost exclusively of boulders and fragments of Elbrus lava and tuffaceous breccia. This fact unquestionably points to a lower Tertiary-upper Apsheronian age of the oldest moraine of the northwestern Elbrus slope. There are no higher terraces and higher moraines in that area. The Bryn-Tash pass, like the vast Irakhik-Syrt plateau east of it, is also made up of the oldest glacial deposits, this time along the right bank of the Malka.

Typical well-stratified glacial outwash deposits are developed below the Ullu-Chiran glacier, along the left side of the canyon. It is to be noted that the elevation of the western part of the Irakhik-Syrt plateau, which is an eroded ground moraine of the oldest Elbrus glacier, is only slightly less than that of the Bryn-Tash pass (3067 m), with the depth of the Malka cut about 200 m. This also corresponds to the relative height of the Upper Apsheronian (Gunz) terrace.

S. P. Solov'yev [27, 28] notes numerous traces of ancient glaciation at different elevations (up to 600 m along the left side of the Baksan valley), in canyons of Azau, Garabash, Terskol, Irik, Chat, and Kyrtyk rivers. A large terminal moraine, making up the Tyubele hill (1762 m), is located 15 km downstream from the confluence of the Azau and Donguz-Orun, on the left bank of the Baksan, just below the mouth of the Kubasantych.

There are no terminal moraines, or any moraine deposits at all, down the Baksan valley. The alleged moraine deposits mentioned in the literature, below that point, all the way to Zayukovo (65 to 70 km distant, elevation about 170 m), turned out to be outwash deposits or else alluvial cones of the tributary canyons. The upper (Apsheronian) terrace at Zayukovo does not "hang in the air," as stated by some authors (Yu. P. Masurenkov and others) but rather enters the Baksan canyon. The relative elevation of the Tyubele moraine is about 180 m; down the valley, there are terrace remnants of about the same elevation. This makes it reasonable to assume a Gunz (upper Apsheronian) age for that most ancient of the moraines.

Ancient glaciation was also developed along the west Elbrus slope. Thick moraine deposits have been observed up to elevation 1600 m, along both upper tributaries of the Ullu-Khurzuk: Biy-Tyk-Tyubyu (right) and Kyukurtlyu (left). They have been greatly eroded, downstream, while clean-cut terraces, approximate elevation 150 to 200 m, are seen along the Ullu-Khurzuk. Along the Ullu-Kam valley, an ancient glacier reached 10 km below the confluence of its headwaters, forming a well-defined terminal moraine (elevation above sea level, 1800 m), above the Uzun-Kol canyon.

The ancient snow line descended to about 2800 m [35], only some 600 to 700 m below the present line.

M. V. Muratov and M. V. Gzovskiy [19] recognize only two glaciations (Riss and Würm) for the northern Elbrus slope, while Ye. Ye. Milanovskiy [15] believes there is evidence of three glaciations in the Elbrus area (Riss and two Würm periods).

According to our own observations, the moraine overlying the Malka flow lavas is correlative with the ancient moraine of the Irakhik-Syrt plateau. As such, it cannot be Würm. On the basis of the above evidence, it is lower Tertiary (upper Apsheronian).

CONCLUSIONS

The geologic history of the area is very complex, involving as it does rocks from the

Precambrian to and including the upper Tertiary, with greatly disturbed Paleozoic rocks.

Structurally, the subject area comprises three tectonic zones: 1) central (block) uplift of the Main Range; 2) block-folded zone of the Front Range; and 3) the North-Caucasian monocline, which determines its geologic features.

With relation to the first zone, the volcanogenic Elbrus rocks comprise the so-called Main Range granite, intruded into crystalline schist. The second tectonic zone is made up of assorted sedimentary and volcanic rocks, Lower to Middle Devonian and Carboniferous in age, with a small development of lower Permian and Liassic rocks. This zone represents the northern wing of a complex major anticline with a stratigraphic sequence from older rocks in the south to younger, in the north. Secondary folds are present; faults are mostly normal. No transverse (meridional) disturbances have been found. The third (northern) tectonic zone is made up of middle and upper Liassic rocks (Toarcian).

In the Precambrian (?), the subject area was part of a vast geosyncline with an accumulation of thick sedimentary and volcanic rocks. They were altered by subsequent regional and contact metamorphism to assorted types of crystalline schist.

A subsidence took place in the Lower and Middle Devonian, with an accumulation of thick marine deposits, and intensive volcanic activity as a result of which the rocks were dynamically metamorphosed.

Prior to the upper Carboniferous, the "Main Range granite" ("gray," of the Balkar type) was intruded and the region underwent a radical change in its geosynclinal environment. The middle and upper Carboniferous were times of continental sedimentation. There was a break in sedimentation, prior to the lower Permian. Sediments of that period were accumulated under continental-lagoonal conditions, with some volcanics accumulating. After another break, thick multicolored conglomerate was deposited.

The region appears to have stood high, in the upper Permian and the Triassic. Only after the ancient Kimmeridgian folding phase was its northern rim involved in a subsidence, with Toarcian sediments accumulating in a Molasse sea. In later time, the region underwest emergence and erosion, with a revival of volcanism in the Oligocene (Maikop), as a sequel to a pre-Oligocene orogenic phase. As a result, a thick (over 2 km) volcanic sequence was accumulated on the ancient erosional surface.

Only block displacements of various magnitudes could take place in this rigid framework in the Neogene when Alpine type relief was initiated. Block displacements occurred here, during the pre-Akchagyl orogenic phase, with the Elbrus volcano originating at an intersection of old faults, in the central, the most elevated area.

The Quaternary witnessed a development of glaciation and of intensive river erosion which was uneven in different zones, because of the persistent movement along ancient fault planes. A study of the longitudinal and transverse cross-sections of the upper Malka valley, within the block-bolded Front Range zone (Fig. 5), shows that the blocks were differentially uplifted, with the middle block relatively the highest. All blocks are tilted to the north; this is probably because of the general dome-like uplift of the Main Range province.

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Received July 6, 1958

NEW DATA ON POST-JURASSIC MAGMATISM OF NORTHWESTERN CAUCASUS[1]

by

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Despite the extended earlier geologic study of the Caucasus, a more detailed acquaintance with its individual regions brings forth new data valuable for both general petrology and a better understanding of the geologic history of this folded province.

Prior to 1955, the western terminal of the Main Caucasian Range (upper courses Pshish, Psekha, Tuapsinka rivers) was believed to be an area of development of Jurassic flysh (Goytkh formation of O. S. Vyalov) and subordinated quartz porphyry tuff and tuffite (Indyuk Mountain formation). In 1955, G. D. Afanas'yev [1] demonstrated that the Indyuk Mountain rocks consist of peculiar "extrusive" bodies of a granitoid composition, which cut the Jurassic flysh and are younger than enclosing Toarcian-Aalenian shale.

Our subsequent work in this densely wooded area has disclosed a wide development of secondary trachytic rocks. At the same time, we found exposures of peculiar types of breccia, similar to those of the Two Brothers Mountain and other places. Brief communications on these discoveries were published in 1957, 1958 [3, 4].

This paper presents the latest data on the geology of the western terminal of the Main Range and on specific cases of magmatism, chiefly Cenozoic. The new data on the latter are so important for the geology of the Northern Caucasus that we deem it expedient to publish some of them.

A. BRIEF GEOLOGIC SKETCH

Earlier study of this part of the Caucasus by O. S. Vyalov [13, 14], V. V. Belousov and B. M. Troshikhin [7], G. D. Afanas'yev and Borsuk [4], together with recent investigations, has disclosed the following sedimentary and igneous section in the western part of the Main Caucasian Range (Fig. 1) [reading upward].

1. Lower Jurassic shale with sandstone and siderite beds (Goytkh formation of O. S. Vyalov).

2. Diabase porphyry, Maloye Pseushko area.

3. The Altubinal Lower Cretaceous (?) tuffaceous sediments.

4. Intrusive granodiorite porphyry (quartz porphyry) approximately contemporaneous with the Altubinal sequence.

5. Extrusive granitoids, of the Indyuk Mountain type; their absolute age, as determined on numerous occasions, is 90 to 110 million years, which makes them, most likely, Upper Cretaceous.

6. Subalkaline gabbroids (essexite, crinanite, teschenite?).

7. Eruptive breccia of the Two Brothers type, and carrying the material of older rocks.

8. Eruptive breccia of the southern slope of Semashkho Mountain, supposedly carrying rock fragments from the preceding group.

9. Extrusion of sodium-rich porphyry near Krivenkovskaya station.

10. The "watershed" eruptive breccia of trachyte tuff, fragments of earlier rocks, and Cenozoic-type trachyte (bostonite). The absolute age of the trachyte has been determined to be as much as 30 million years, corresponding to the Eocene or Miocene.

This section does not include sandstone with argillaceous cement, occurring as placers on many watersheds. There is

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[1]Novyye dannyye o posleyurskom magmatizme Severo-Zapadnogo Kavkaza.

evidence that they are nearly contemporaneous with the watershed breccia.

1. The lack of space prevents a detailed description of Lower Jurassic sedimentary deposits. The stratigraphy, lithology, and tectonics of this section have been treated elsewhere [7, 12, 13, 14].

Here we only note that Lower Jurassic flysh of the subject area has been folded into a system of complex flexures, trending northwest, with the Indyuk Mountain type "granitoid" extrusions associated with their anticlinal parts. Judging from the change in the position of Lower Jurassic shale in the western part of the area (Krivenkovskaya station), which assume a northeastern strike there, it may be supposed that the anticlinal axis of the Main Ridge flysch plunges southeast, toward the headwaters of the Pshekhi.

A dome-like extrusion of sodium-rich porphyry at Krivenkovskaya station, discovered in 1958, and the tabular, northeast-trending "granitoid" extrusions of the Indyuk Mountain type, gravitate toward the western part of the structure, toward its rise. The "granitoids" are described in a paper by G.D. Afanas'yev, 1955 [1].

Younger formations -- the Altubinal and others -- have been preserved for the most part in the eastern segment of the anticlinal structure.

2. Diabase porphyrite, exposed at Maloye Pseushkho, is the oldest magmatic rock of the area. It forms dikes which cut Toarcian shale, and is exposed in a flood channel of the Pshiyakho, at the southeastern edge of Maloye Pseushkho.

The age of the diabase porphyrite is post-Toarcian, since it definitely cuts the Toarcian flysch. Boulders of petrographically similar diabase-porphyrite have been found in Lower Cretaceous (Valanginian, according to O.S. Vyalov) arkosic sandstone alternating with limestone, in the Maloye Pseushkho Mountain.

3. The Altubinal tuffaceous sedimentary deposits form steep folds, trending northwest and resting upon an uneven surface of Lower Jurassic shale, of a low mound relief. The largest exposures of the Altubinal sequence, spared by erosion, have been found in the eastern part of the area, in the headwaters of Malyy Pshish.

A thick, highly disturbed sequence of alternating shale and tuffaceous breccia has been found at the headwaters of a left tributary of Malyy Pshish, which flows from the vicinity of the sawmill, and along a right tributary of the Pshish, which joins it above the Altubinal. This sequence rests with an obvious unconformity upon Jurassic shale (marked with the hammer, in Fig. 2). Here, the dislocated volcanics envelop a sharp ridge of Lower Jurassic shale, standing on edge.

The bulk of the tuff breccia consists of fragments of plagioclase, quartz, biotite scales and occasional grains of K-Na-feldspar. It also contains a large amount of red to pink almandine garnet. The shape of breccia fragments is mostly rounded or pear-like, reminiscent of volcanic bombs, with angular, asymmetic fragments occurring alongside the perfectly spherical ones. In the character of their component rocks, the fragments may be divided into the following five groups:

1) Rocks of a sedimentary origin (metamorphic limestone carrying a fossil fauna), sandstone, shale and mudstone, some of which are Paleozoic.

2) Paleozoic microcline granodiorite and plagiogneiss, remnants of an ancient crystalline basement, deeply submerged in the Tuapsinsk region. These remnants were carried to the surface as components of the tuffaceous breccia during the volcanic process responsible for the formation of the Altubinal sequence.

3) Diabase porphyrite, similar to that at Maloye Pseushkho.

4) Volcanic bombs of a dacite type, similar in many respects to granodiorite porphyry (quartz porphyry) of Semashkho, Two Brothers, Lysaya, and Shessi mountains. Dacite fragments (extrusive facies of granodiorite porphyry) in the Altubinal breccia are rounded or pear-like, commonly with an elongated end, and attain large sizes (as much as 0.5 x 1 m). Smaller fragments predominate, however (15 to 20 cm in length). Light-colored extrusives are keratophyric, considerably altered by carbonatization.

4 and 5) Intrusions of granodiorite porphyry and the Indyuk Mountain type granitoid are stock-like bodies of granodiorite porphyry usually associated spatially with the Indyuk Mountain type granitoid "extrusions", and forming two sub-parallel belts with them trending NW, to come together in the Pshekhi headwaters (Shessi Mountain and other areas). They form the crests of Shessi, Lysaya, Two Brothers, and Semashkho mountains. Without pausing for a description of these rocks,. which has been done elsewhere [4], we note here only their most important features.

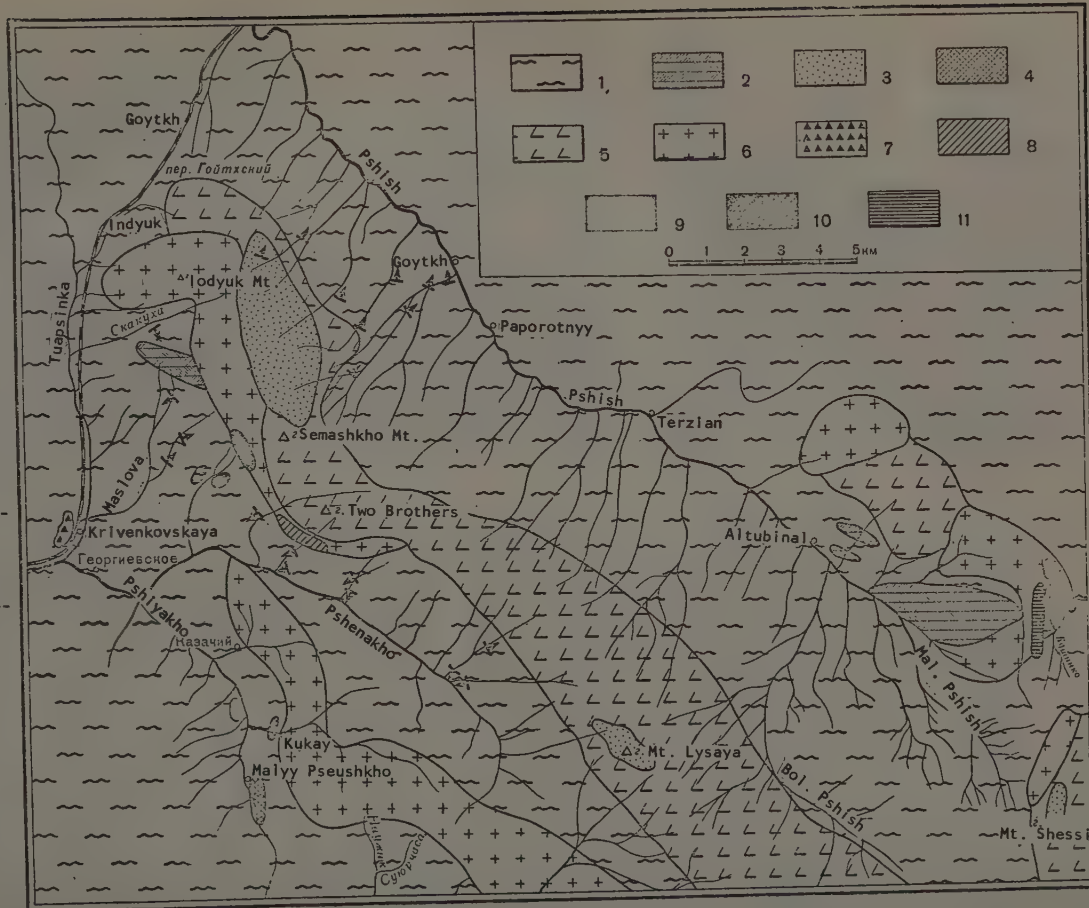

FIGURE 1. Generalized geologic map of the Pshekha-Tuapsinka watershed area of the Main Caucasian Range.

1 -- Jurassic flysch; 2 -- Altubinal volcanic-sedimentary deposits; 3 -- sandstone; 4 -- diabase porphyrite; 5 -- quartz porphyry; 6 -- "extrusions" of the Indyuk Mountain-type granitoid; 7 -- extrusions of sodium-rich porphyry; 8 -- eruptive breccia of the Two Brothers Mountain type; 9 -- pyroxene gabbro; 10 -- eruptive breccia of the Semashkho Mountain; 11 -- watershed eruptive breccia.

FIGURE 2. Volcanogenic sedimentary Altubinal sequence unconformable over disturbed Lower Jurassic sediments (right bank ravine of Malyy Pshish above Altubinal). The volcanics envelop a denuded remnant of folded shale.

There is no doubt but that granodiorite porphyry cuts the Toarcian-Aalenian flysch. It contains xenoliths of diabase porphyry, petrographically similar to those occurring in the Pseushkho area. The structural differences between the Two Brothers Mountain granodiorite porphyry and the dikes confirm its intrusive nature, relative to Lower Jurassic flysch. This is also suggested by several sharp unconformities and the granodiorite porphyry cutting relationship to the shale.

Thus, the position of the granodiorite porphyry and its structural features contradict the accepted idea of its being a blanket formed synchronously with the Jurassic flysch.

New data on the Indyuk Mountain-type granitoid intrusions, as compared with the old ones [1, 2, 3, 4], point out new evidence of their being younger than the Altubinal formation. The intrusions cut this formation, assimilating and granitizing the contacting clastic beds. It has also been shown that granitoids are accompanied by their vein facies -- aplitic rocks with a structure and composition of their own. Their intrusive mechanism is different from the standard for that of granite magma. It has been considered to some extent, in our earlier papers. More complete data are to be presented in a forthcoming work.

6. Gabbroid rocks of a higher alkalinity are exposed in many localities of the subject area, in dikes of various sizes. These outcrops are everywhere concentrated around zones of eruptive breccia of the Two Brothers Mountain type. The most important are: 1) southern slope of the Main Range, in the Semashkho and Two Brothers mountains; 2) left side of the Malyy Pshish valley, in its upper course, near the sawmill; 3) upper course of the Kushinko valley (left tributary of the Pshekha). The Chernyy Kamen' rock, in the Pshiyakho valley, near Maloye Pseushkho settlement, is made up of similar rocks.

Gabbroid rocks are best developed in the Kushinko basin, where its relationship with other formations of the area is graphically exposed.

Cliffs of extrusive granodiorite of the Indyuk Mountain type rise up along the northeastern slope of the Pshisha-Kushinko watershed (Fig. 3, 4). Kushinko River, in cutting this granitoid, has exposed younger magmatic formations. A large area of the river bank section is occupied by a Two Brothers Mountain type eruptive breccia. Gabbroid rocks appear at the waterfall, where they are represented by the following varieties:

1) Large dikes of a black rock, locally with a well-defined parting; petrographically similar to the so-called "crinanite." They cut a shattered and brecciated formation carrying assorted chunks of Jurassic shale, granitoid, and remnants of rocks petrographically like sub-alkaline gabbroids.

2) Exposed in the lower part of both banks of Kushinko River, there are massive rocks forming a dome-like extrusive body buttress-

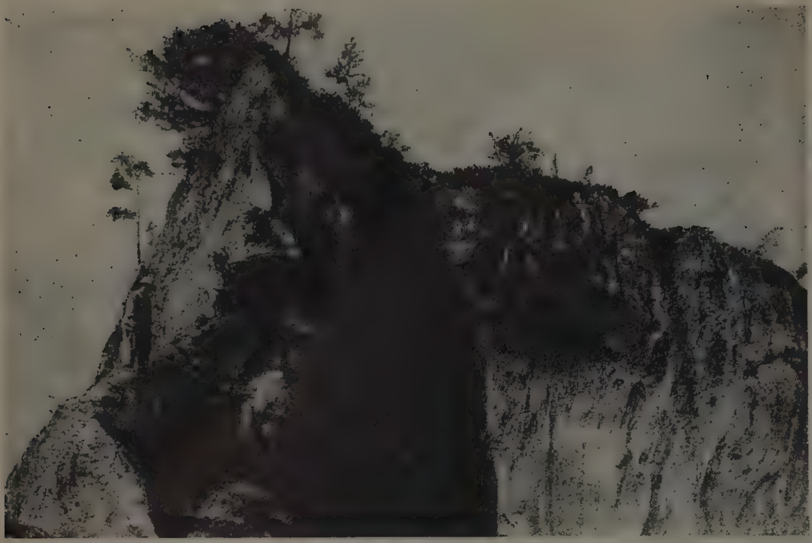

FIGURE 3. Cliffs of the Indyuk Mountain type "extrusive" granitoids, along Kushinko River, as viewed from the west, from the Pshish watershed.

FIGURE 4. Same, as seen from a point 250 m below, at Kushinko River, to the east.

a -- granitoid "extrusion" of the Indyuk Mountains type; pyroxene-free gabbro; b -- neck; c -- vein.

ing the black gabbroids and sending off apophyses of veins and blind stocks into the Indyuk Mountain type extrusive granitoids. "Crinanite" above the extrusive gabbroids is locally uplifted and shattered to fragments, with apophyses of the underlying gabbroid penetrating the shattered zone. These relationships are seen in Fig. 4, 5.

3) The "crinanite"-type black gabbroid contains a vein of aplitic teschenite with sanidine incrustations. The age of these veins has been determined by the fact that they cut first the Lower Jurassic shale, then Upper Cretaceous (?) granitoid extrusions (absolute age of granitoids, 90 to 110 million years).

Eruptive breccia of the Two Brothers Mountain type, which rests upon gabbroids and contains remnants of similar rocks, is for that very reason a component of a single complex which includes them both. On the other hand, gabbroid fragments of the same type, but of a secondary origin, occur in tuffaceous formations and eruptive breccia of the watershed type. Consequently, the age of these gabbroid rocks is not older than Eocene and not younger than Miocene.

7. Eruptive breccia of the Two Brothers Mountain type. East of the breccia exposures on the southern Semashkho Mountain slope, 2 to 3 km from the top, on the left bank of a stream flowing from the Two Brothers, there are outcrops of an eruptive breccia

FIGURE 5. Pyroxene gabbro, buttressed by an extrusion. Zone of eruptive breccia in gabbro, above the extrusion.

1 -- pyroxene gabbro; 2 -- high alkaline gabbro extrusion; 3 -- brecciated gabbro.

appreciably different from that of Semashkho Mountain and very much like that of Kushinko Mountain where it rests at the top of gabbroid extrusions. Eruptive breccia of the Two Brothers area has been formed in the contact zone between Jurassic shale and Cretaceous (absolute age determination) gabbroid extrusions of the Main Range. The zone 60 to 70 m thick between them is represented by a peculiar breccia with its bulk consisting of a black, coarse-grained matrix. Its peripheral parts carry irregularly-shaped and poorly-sorted granitoid fragments of the adjacent types (to the north), and of the Jurassic shale near the contact.

Besides these country rocks, the breccia carries numerous remnants -- generally reminiscent of volcanic bombs, with rarer angular chunks of assorted extrusives. Present among them are the Kushinko River alkaline gabbroid rocks, especially their extrusive representatives, also chunks and fragments of trachytoid rocks, Paleozoic microcline granite, and occasional limestone.

In the composition of its matrix and fragments, the Two Brothers Mountain breccia is quite similar to that found at the top of the Kushinko Mountain gabbroid extrusions.

The broken topography and dense forest of the area allow only an estimate of the form and size of the brecciated body of the Two Brothers Mountain. In a river cut, its thickness across the strike is about 70 m. The breccia extends 200 to 300 m northwest, along the left bank, to the watershed where it is concealed under alluvium. No breccia exposures have been found along the strike of this zone, in a ravine about 1.5 km east of there, or in the sodded right bank of the river.

8. Eruptive breccia of the Semashko southern slope. The following section has been exposed by an unnamed creek along the wooded southern slope of Semashkho Mountain, reading downstream: 1) sedimentary rocks, chiefly Lower Jurassic shale with sandstone beds; 2) extrusive granitoids of the Indyuk Mountain type; 3) a break in the exposure, followed by about 20 m of an eruptive breccia.

Breccia fragments, generally angular, from a few to 30 cm, are represented by sodium trachytoid, felsite of a sodium porphyry type (keratophyre) and rocks petrographically similar to the Kushinko River gabbroid. The breccia cement -- either an extrusive rock of the albitophyre type or a tuff of the same composition -- is cut by analcime and calcite veinlets. Some fragments are strongly zeolitized.

Figure 6 illustrates an eruptive breccia polished section from Semashkho Mountain.

There are exposures of extrusive granitoid, down the ravine, cut by dikes of gabbroid similar to those of the Kushinko River.

Additional work is needed to find out whether these two breccia exposures, on the Two Brothers and Semashkho mountains, are individual formations associated with a single structure, i.e., "explosive vents," or if they are the same formation, separated by faulting.

The field experience with this peculiar breccia, and the composition of the Kushinko

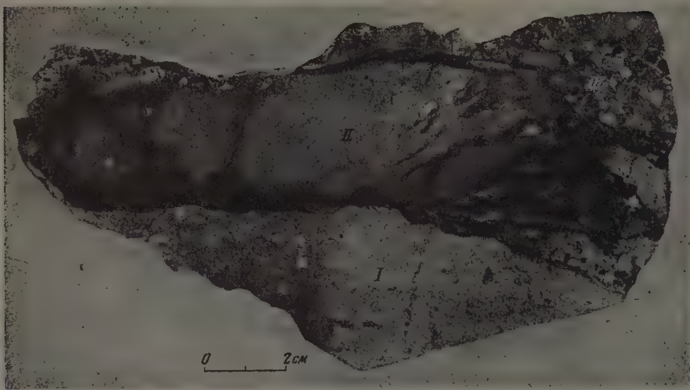

FIGURE 6. Eruptive breccia.

I -- Fragment of sodium trachyte; II -- felsite; III -- cement with analcime and calcite veinlets.

River breccia, favor the first assumption, i.e., the "explosive vent" origin. According to field data along the Kushinko, these eruptive breccias are buttressed from below by intrusive and extrusive rocks of a series which we associate with highly alkaline gabbroids.

9. The sodium-rich porphyry extrusion at Krivenkovskaya, about 200 square m in area and possessing many interesting features, was discovered as recently as 1958, as a result of a washout in the road at Krivenkovskaya. Long rains uncovered a shelf of porphyry in contact with Jurassic shale. The age of this leucocratic sodium-rich porphyry was determined as post-Lower Jurassic because of its obvious intrusion into Lower Jurassic shale. The intrusive body is dome-like, somewhat elongated to the northwest. Its upper age limit is unknown, although there is indirect evidence of its being close to that of the southern Semashkho eruptive breccia. Both formations carry xenoliths of sodium trachyte.

10. Eruptive breccia of the "watershed" type

FIGURE 7. Exposure of the Kushinko basin eruptive breccia.

make up a zone about 100 m thick, extending meridionally along the eastern slope of the Malyy Pshish-Kushinko watershed, over a distance of about 2 km. A zone of similar breccia trends nearly latitudinally in the Malyy Pshish upper course. The enclosing rocks of the "watershed" breccia are compact clay, a product of decomposition of tuffaceous trachytic material.

Along the Kushinko River, this breccia (Fig. 7) borders sandstone on the west (the youngest sediments of the area, next to the topsoil loam), and "extrusive" granitoid on the east.

The peripheral parts of the breccia zone carry remnants of the corresponding sandstone and granitoid with especially numerous chunks of leucocratic trachytoid rocks of the bostonite type. Similar rocks have not been found in the Two Brothers Mountain type breccias.

The present data suggest that these tuffaceous "watershed" breccias are the youngest formations of the region, next to surface loams. They cannot be older than Eocene and are not likely to be younger than Miocene.

B. PRELIMINARY RESULTS OF A PETROGRAPHIC STUDY OF CENOZOIC MAGMATIC FORMATIONS OF THE MAIN CAUCASUS RANGE (BASIN OF PSHEKHA, PSHISH, AND TUAPSINKA RIVERS)

The lack of space precludes a petrographic description of the Altubinal volcanic sedimentary sequence, the quartz porphyry extrusions, and peculiar granitoid "extrusions." Here, we shall only note that assorted extrusives are widely developed in that sequence. Specifically, stratified shale with beds carrying extrusive boulders, and described in a 1957 paper [4], now may be definitely designated as a separate formation (the Altubinal volcanic sedimentary formation).

Kushinko River Gabbroid

As pointed out before, granitoid intrusions and stratified Lower Jurassic shale have been cut by highly alkaline gabbroid, in the upper course of Kushinko River and in a number of other localities. Pictures in Figs. 3, 6, and 7 illustrate the relationship of magmatic rocks in the Kushinko River area.

The granitoids have been cut by apophyses shooting off from olivine-free sub-alkaline gabbroid extrusions which are exposed in the Kushinko channel. These extrusions acted upon earlier stocks and dikes of augite feldspathic gabbroids, with the resulting formation of eruptive breccias at the top of the extrusions. Gabbroids are cut by some thin veins of an aplitic rock, melanocratic in appearance but containing virtually no dark minerals.

Gabbroid veins (essexite gabbro, crinanite ?) consist of dense, massive, crystalline rocks, dark to almost black, with plagioclase and pyroxene inclusions perceptible to the naked eye. They are porphyritic with a gabbroid matrix, consisting of plagioclase, pink to lilac monoclinic pyroxene (titanium-augite), with a small quantity of biotite tablets, and olivine recognizable by its form but replaced by chlorite-serpentine and a carbonate. A considerable development of zeolites (scolecite), mostly on plagioclase, occurred during the last crystallization stage.

The intergranular matrix space contains zeolite (scolecite?) and possibly nepheline (a low birefringence mineral, with a refractive index of 1.53 to 1.54, and direct extinction). Chlorite and carbonate, with accessory apatite are newly-formed minerals; magnetite and ilmenite are ore minerals which are present.

Plagioclase accounts for as much as 60% of the rock. The first generation of plagioclase, represented by bytownite and corresponding to an intratelluric phase, forms large (4.8 to 5.2 mm) idiomorphic phenocrysts of a widely tabular or elongated prismatic habit. The orientation measurements of the optical indicatrix on Fedorov's table, as well as in section ⊥ (100), show that the central part of the large incrustations corresponds to Nos. 62 to 70. An interesting phenomenon is a common and considerable replacement of the central part of the first generation plagioclase inclusions, by zeolite. The altered core of plagioclase becomes overgrown with more acid plagioclase (labradorite No. 60). The second generation grains are of the same composition; they make up the bulk of the matrix. Zeolites which are developed on the central parts of larger plagioclase inclusions are transparent to passing light, have a low refractive index, and a birefringence of about 0.008, 2 V = -58 (along two axes of the Fedorov table). This zeolite is close to scolecite in its optical properties.

Monoclinic pyroxene -- titanium-augite accounting for about 20% of the rock -- is similarly present in two generations. The first generation pyroxene, whose precipitation from the melt took place at the same time with that of first generation plagioclase (intratelluric phase), formed large idiomorphic crystals, less intensely colored. In transmitted light, pyroxene has a weak pink to

Table 1

Chemical composition of Cenozoic magmatic rocks of the Western Caucasus

Specimen No.	Rock and locality	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	CO_2	Loss on ign.
	Alkaline Gabbroids of Kushinko and Pshish Rivers and the Two Brothers Mt.												
481-e/57	Topmost gabbro of Pshish River	39.02	1,75	17,67	4,28	6,00	0,65	3,84	13,06	2,82	0,18	5,80	3,02
54/58	Same for Kushinko River	46,42	—	—	—	—	—	—	—	2,30	0,78	0,84	4,02
396/58	Gabbro extrusion	44,94	—	—	—	—	—	—	—	3,04	1,45	1,94	5,31
54-z/58	Gabbro extrusion	45,35	—	—	—	—	—	—	—	2,89	1,34	1,60	4,80
318-a	Vein gabbro of the Two Brothers Mt.	—	—	—	—	—	—	—	—	2,00	0,90	—	—
391/56	Vein gabbro of the Pshish River	—	—	—	—	—	—	—	—	2,60	0,85	—	—
401/58	Vein aplite of teschenite	45,93	—	—	—	—	—	—	—	0,44	4,20	11,8	2,71
	Eruptive Breccia of the Two Brothers Mountain Type												
315/56	Alkaline trachyte	52,74	—	—	—	—	—	—	—	1,02	10,82	7,94	—
316/56	Vitrophyre of a basic rock	43,60	—	—	—	—	—	—	—	3,42	1,75	—	—
166-d 55	Trachyte	54,88	—	—	—	—	—	—	—	2,36	7,43	—	—
318/55	Alkaline trachyte	—	—	—	—	—	—	—	—	0,37	11,13	—	—
607-b—x/56	Leucocratic trachyte	63,90	0,37	16,20	1,76	0,56	0,08	0,31	2 78	4,92	6,50	2,20	0,58
607-a—x/56	Leucocratic trachyte	68,60	0,30	15,00	1,76	0,56	0,03	0,21	2,02	7,14	3,27	0,88	0,41
405/56	Vitrophyric variety of crinanite	47,02	—	—	—	—	—	—	—	3,43	3,98	6,60	—
394/56	Alkaline trachyte	50,10	0,32	14,28	0,64	0,78	0,98	0,39	11,78	4,83	5,04	2,76	1,28
373/56	Alkaline trachyte	—	—	—	—	—	—	—	—	4,29	5,39	—	—
	Eruptive Breccia of the Semashkho Southern Slope												
28-n/58	Fragment of sodium-rich trachyte	69,01	—	—	—	—	—	—	—	8,53	0,18	1,26	0,65
32/58	Breccia cement--sodium-rich trach. (w/sec. quartz)	70,17	—	—	—	—	—	—	—	6,97	0,08	0,60	0,65
102/58	Extrusive porphyry at Krivenkovskaya	60,74	—	—	—	—	—	—	—	6,44	0,05	9,16	—
128/57	Kukay River gabbroid dike	44,64	0,84	15,01	1,30	9,11	0,65	3,89	6,38	5,65	0,28	8,80	3,11
	Watershed Breccia												
225-a/57	Trachyte (bostonite) from breccia lumps	67,07	—	—	—	—	—	—	—	0,14	14,53	—	—
225/57	Clay, decomposition product of trachyte tuff, (breccia enclosing rock)	61,02	0,57	17,21	4,48	1,55	0,05	1,98	0,04	1,02	4,52	0,30	4,82

Note: Comma represents decimal point.

lilac coloring, barely perceptible pleochroism, $2V = +50^o$, a 50^o angle with γ, $\gamma - \alpha = 0.025$. Second generation pyroxene, of a deeper pink-lilac, is represented by fine xenomorphic grains.

Deep brown, highly ferruginous biotite is present only in small amounts, as small tablets in the matrix and as inclusions in the incrustations.

Olivine is assumed to be present, because of the presence of pseudomorphs of carbonate and chlorite serpentine on prismatic grains of irregular form, which appear to have been associated originally with this mineral.

Zeolite of the scolecite type partly replaces the central part of plagioclase grains and is also formed in the matrix interstices.

In their chemical (Table 1), and mineral composition and geologic position, these rocks are closest to alkaline gabbroids of the crinanite (?) type, or to certain types of olivine teschenite (?).

Sub-Alkaline Extrusive Gabbroids (Kushinko)

These are dense porphyritic rocks, gray-black with a greenish tint, with a cryptocrystalline matrix containing plagioclase incrustations, 1.5 x 1 mm, accounting for 16% of the whole. The matrix, accounting for 84%, consists of plagioclase microlites and ilmenite tablets in a cryptocrystalline mesostasis.

Plagioclase phenocrysts have a coarse tabular and elongate prismatic habit. In its composition, plagioclase corresponds to labradorite No. 60. It lacks zonation. There is a strong pseudomorphism of low-refringence zeolite of the scolecite type, with a negative angle of optic axes, of 57 to 76^o, and a birefringence on the order of 0.008 to 0.010. Its refractive indices for a thin section are $\gamma = 1.519$ and $\alpha = 1.511$.

These rocks carry a small amount of biotite and commonly contain rounded vugs with zeolite and a carbonate; their texture is porphyritic, with a cryptocrystalline matrix.

The lack of complete chemical analyses of these rocks prevents their exact placing in the alkaline gabbroid group. It is unquestionable, however, that they are undersaturated with silica (Table 1) and contain more alkali, than the preceding rock group, for the same silica content.

Kushinko River Teschenite Aplite

The Kushinko gabbroids are cut by a vein of dark fine-grained rock 8 to 10 cm thick consisting chiefly of an unidentifiable, very finely crystalline substance with a refractive index just below that of Canada balsam. There are some phenocrysts of a perfectly transparent sanidine, with isometric outline and optical constants $2V = -48^o$, indicatrix orientation relative to $\perp(100)$ - $\gamma = 90^o$, $\beta = 5^o$; $\alpha = 88^o$. The rock is considerably carbonatized (Table 1, Spec. 401/58), with the carbonate forming aggregates in the

Table 2

Optical properties of plagioclase and non-contemporaneous basic rocks of the Tuapsinsk area

Specimen No.	Rock and Location	Plagio-clase No.	2V	Orientation of optical indicatrix			Twinning Law
				γ	β	α	
428/56	Diabase porphyrite, Maloye Pseushko area	45	+88	86	62	27	Albite
428-s/56	Same	40	+86	85	71	20	"
428-L/56	"	55	—	85	79	12	Baveno, right
64/57	"	50	—	45	55	65	Carlsbad
481-e/56	Gabbroid, Pshish and Kushinko headwaters, and the Two Brothers area	85	+88	46	60	58	Carlsbad
481-b/56	Edge zone	68	+78	68	62	36	Albite
33/57	Gabbroid, Kushinko River	82	-84	62	60	43	"
33-a/57	Same	78	—	67	31	71	Albite-carlsbad
54-zh/58	"	90	-88	73	32	64	" "

Table 3

Optical properties of K-Na-feldspar in the Tuapsinsk area trachytoids

Specimen No.	Rock and Location	2V	Optical orientation		
			γ	β	α
166-d/55	Trachyte fragment in the Two Brothers Mountain type breccia	-48	82	10	82
315/55	Same	-54	—	—	—
399/55	"	-60	88	3	88
20/55	"	-48	86	12	78
605-a/56	"	-62	—	—	—
605-x—zh/56	"	-50	89	21	80
605-x—b/56	"	-32	—	—	—
405/56	Malyy Pshish trachyte, near sawmill	-58	88	18	72
605-z/56	Sanidine in matrix of the Two Brothers-type breccia	-12	—	—	—
605-x/56	Same	-24	—	—	—
151-i/57	Trachyte fragment in "watershed" breccia	-48	90	12	77
225-zh/57	Same	-50	87	8	83
115-e/57	"	-60	86	8	82
225-a/56	"	-24	86	5	90
100-a/58	Sanidine fragments in "watershed" breccia	-38	85	11	80

matrix and partly replacing the sanidine. Occasional biotite grains are the dark-colored minerals and the ore minerals are chalcopyrite, and pyrite.

Table 1 lists new data on the chemical composition of rocks for the northwestern Caucasus, while Tables 2 and 3 give results of optical measurements of feldspars of some of the groups.

Eruptive Breccia Rocks of the Two Brothers Mountain Type

The matrix of the Two Brothers Mountain breccia is a dark rock of individual water-clear quartz crystals and less common small fragments of extrusives and crystals of a dark-red garnet. Also present are sulfides in spots, native copper and occasional sodium-rich white zeolite. The composition is more or less homogeneous throughout the breccia. Its black color is probably due to pulverized carbonaceous matter from ground Lower Jurassic shale present as chunks in the breccia.

The extrusive fragments in breccia are of alkaline trachyte, a light-gray dense rock with rare incrustations of a sanidine-like anorthoclase in the matrix of elongated prismatic microlites of the same anorthoclase, floating in a cryptocrystalline substratum of radiated aggregates with a K-Na-feldspar structure (Table 1, spec. 315/56 and 318/55). The high alkaline content (chiefly K_2O) and low silica content (52%) lead to the belief that isolated tablets of a low birefringence mineral with n larger than that of K-Na-feldspar, observed in thin sections, are of nepheline (?). Ore minerals are represented by pyrite and chalcopyrite, and there are abundant pseudomorphs of calcite on feldspar.

Leucocratic trachyte is light-colored, feldspar-rich rocks with dark minerals nearly lacking (Table 1, spec. 607-a—x/56), but with secondary quartz in vugs.

Vitrocrinanite is a dark-gray dense rock consisting of a brown vitreous matrix with concretions of ore minerals and incrustations of basic plagioclase in the process of replacement by zeolites. A partial analysis (Table 1, spec. 316/55) shows its similarity, in the alkaline and SiO_2 content, with the

Kushinko River gabbroids (spec. 396/58).

Chunks and fragments of alkaline leucocratic trachyte (Table 1, spec. 394/56) and of a vitrophyric alkaline gabbroid (spec. 405/56) have been found in bluish clay, a product of decomposition of the tuffaceous matrix of the Pshish River breccia (at the sawmill location).

Rocks of Eruptive Breccia of the Semashkho Mountain Southern Slope

The Semashkho Mountain breccia consists essentially of extrusive fragments. The amount of cement is comparatively small. Widely developed in the fragments are light-colored sodium-rich crystalline trachytoids, virtually lacking in potassium. Also present are large, somewhat rounded fragments of altered gabbroid of the Kushinko River type.

The breccia cement is a sodium-rich trachyte-type extrusive. Externally, it is a dark-green porphyritic rock, with a chiefly crypto-crystalline matrix carrying plagioclase incrustations, and with fractures and vugs filled with carbonate and analcite. There are occasional grains of secondary quartz in vugs.

The matrix, accounting for 74% of the rock, consists of microlites of acid plagioclase, i.e., albite-oligoclase. Its texture generally is typically trachytic.

Incrustations are represented by large (2.0 x 1.2 mm) elongated, prismatic to wide, tabular albite grains, in simple twins according to the albite law, with refractive indices $\gamma = 1.530$ and $\alpha = 1.526$. Albite incrustations account for 19.3% of the bulk, with the albite appreciably pelitized.

Among epi-magmatic minerals is quartz, 6.7% of the total. In association with carbonates, analcime fills up cavities in cement and in some breccia fragments. Pyrite is present in minute specks scattered throughout the matrix, as well in some well-formed crystals.

The cement is cut by veins of calcite and zeolites, with analcime predominating, in regular crystals with refractive index 1.483. Locally, the cement is represented by tuff of the same composition (from the sodium-rich trachyte fragments).

A partial chemical analysis of the extrusive cement (Table 1, spec. No. 32/58) has confirmed the high sodium content of the cement. The SiO_2 content, excessively high for a trachyte rock, is explained by the presence of over 6% of secondary quartz.

The sodium trachyte fragments represent a rock externally light-gray, almost white, of a porphyritic texture, with a cryptocrystalline matrix and albite No. 3 phenocrysts (2V = +88). The matrix accounts for 77% of the rock and consist of extremely fine microlite spicules of albite, enveloping the albite phenocrysts. The rock is trachytoid. The albite incrustations account for 17.6% of the bulk, with up to 5.4% secondary quartz in vugs.

A partial chemical analysis shows a calcium enrichment of the rock and its essentially albite character (Table 1, spec. 28n/58).

The Krivenkovskaya Intrusion of Sodium-rich Porphyry

Externally, this is a dense white rock with small incrustations. Under the microscope, the matrix appears to be finely crystallized, essentially feldspathic, with 0.4 x 0.2 mm albite No. 5 incrustations with positive $2V = 84^{\circ}$, and fine quartz grains. The rock is in the process of carbonatization and carries a considerable amount of fine-grained, ore-forming accessory minerals: arsenopyrite, molybdenite, chalcopyrite, and occasional topaz.

A partial chemical analysis has revealed the sodium-rich nature of this near-surface, finely-crystalline intrusive rock. Its geologic relation with the sodium-rich rocks of the Semashkho Mountain breccia is not clear. Considering, however, that "sodium-rich trachyte" xenoliths occur in the Krivenkovskaya intrusion, it is reasonable to suppose that intrusions of this type represent deep-seated portions of volcanic hearths responsible for the Semashko Mountain-type eruptive breccias.

The same group of sodium-rich rocks includes also a dike of light-colored, Lower Jurassic shale poor in SiO_2 along the Kukay (a tributary of Pshiyakho). They are cryptocrystalline, vitrophyric and belong most likely to a leucocratic branch of alkaline gabbroids.

Rocks of the "Watershed" Eruptive Breccia

The matrix of these breccias, unlike that of the Two Brothers and Semashkho Mountain breccias, is represented usually by tuffaceous material of trachyte extrusives. These rocks are in the process of montmorillonization. Externally, they are stone-like clay, dark-gray, weathering to "blue" clay. Fig. 7 illustrates an exposure along the Kushinko River.

Table 1 gives a chemical analysis of dense clay from the "watershed" breccia matrix, similar to that of trachytic rocks, except for a considerable decrease in K_2O content and an increase in water in the clay.

The assorted extrusive fragments represent varieties of basic, trachytoid and other rocks, rich in alkalis, occurring both in outcrops and in fragments of the above-described eruptive breccia. Occurring alongside them in these breccias are numerous chunks and rounded fragments of white crystalline trachytoid rocks, more properly bostonite. They are virtually monomineral, consisting of elongated, prismatic crystals of sanidine-like anorthoclase.

A partial chemical analysis of this rock (Table 1, spec. 225-a/57) confirms its monomineral nature (SiO_2 = 67.07°; K_2O = 14.53%; Na_2O = 0.14%).

It is of interest that attempts at the determination of the absolute age of these potassium-rich rocks have been successful only in isolated instances. A number of samples have a high content of argon. Absolute age figures for different samples range from 35 to 60 million years, with 70 million years for clay, the weathering product of tuffaceous material from specimen 225. This is quite understandable, in view of the "aging" effect of terrigenous material.

The relative geologic position of the watershed breccias, together with the absolute age data, fixes their age as not older than Miocene.

CONCLUSIONS

This petrographic description of the northwestern Caucasus Cenozoic rocks outlines some of their associations.

1. High-alkaline gabbroids of the essexite type. Representatives of this group have certain characteristics of chemical and mineralogic composition in common with the so-called Tertiary crinanite of the islands of Mull and Skye, Scotland and with certain teschenite rocks of Guria (olivine teschenite of Varnet and Okhero, described by D.S. Belyankin).

These rocks are marked by the development of CaO-rich zeolites, during the last stages of crystallization; this is also seen from their high water content (Table 1), as great as 5% in fresh rocks. Carbon dioxide, in these cases, has been determined directly. Rocks of this group occur in sub-intrusive stocks, domes and dikes.

2. Eruptive breccia making up bodies of the explosive vent type. Besides the remnants of country rocks, they contain vitreous varieties of high-alkaline gabbroids and of primarily potassium-rich trachytoids. Data on the Kushinko Mountain gabbroids demonstrate that such eruptive breccia occurs at the top of sub-intrusive gabbroid bodies of high alkalinity.

3. Eruptive breccias of Semashkho Mountain, whose cement and most of whose fragments are essentially (and almost exclusively) sodium-bearing trachyte-type rocks. They are characterized by an abundance of secondary analcime (sodium-rich zeolite), with carbonates in veins cutting the breccia cement and in cavities.

The same high-sodium alkalinity is characteristic of the Krivenkovskaya subintrusive granitoid body (leucocratic Na-porphyry) thus suggesting its age and genetic kinship with the Semashkho Mountain eruptive breccia.

4. Finally, the watershed-type breccia has been described as the youngest magmatic formation of the western Caucasus. Its basement consists of altered trachytic tuff and clay. This breccia carries assorted extrusives, the most abundant -- to the point of exclusion of all others -- being potassium-rich trachytoids, of the bostonite type. The latter are unknown in the older formations of the area.

These preliminary results of the study of post-Jurassic magmatism of the western Main Caucasus Range, and the basins of the Pshish, Pshekha and Tuapsinka rivers reveal a complex picture of its development. The scarcity of exposures, dense forests, and the cover of Lower Jurassic flysh, prevented earlier investigators from deciphering, in the course of routine geologic work, the diversity and details of regional magmatism. At the same time, the area has emerged as very interesting, both petrographically and as an aid in understanding the geologic history and evolution of magmatism of the Caucasus as a whole.

The group of alkaline gabbroids, represented by an association of diversified petrographic types, is closely related to alkaline nepheline rocks proper. The Caucasian data, including those on the Armenian and Georgian rocks, undoubtedly will shed new light on the problem of origin of alkaline rocks.

The problem of the multiplicity of basalt magmas, and the separation of their two types (tholeiite and olivine-basalt), treated in papers by W. Kennedy [18] and J. Verhoogen and F. Turner [20], and their relationship to definite tectonic conditions, also

may be clarified by a study of Caucasian data.

The study of petrographic provinces and epochs, too, will be helped along with new data from further study of Cenozoic magmatism of the Caucasus. It is not accidental that post-Eocene gabbroids and trachytes have been developed in the northwestern Caucasus. Similar associations of Tertiary extrusives are known from other provinces of the Greater Caucasus, from the islands of Mull, Skye and others, from works of A. Harker [17] and G. Tyrrell [21]. A number of papers by I. V. Belov [5, 6] contain interesting material on similar rock associations in Eastern Siberia.

Data on the relationship of different eruptive breccias with the corresponding magmatic rocks, in the specific environment of the northwestern Caucasus, require more study of the formation mechanism of granitoid and gabbroid intrusions and extrusions and its effect on the concentration of useful minerals genetically connected with individual magmas.

Zeolite rocks with a tendency for an alkaline type of association are rare in the Northern Caucasus proper. But analcime-bearing diabase is known from the vicinity of Tubenel', Central Caucasus [16]. A.P. Lebedev's monograph on Jurassic volcanics of the Central Caucasus cites data on diabase of the Tepli and Aday-Khokh ranges. Some of those rocks are similar to gabbroids of the northwestern Caucasus.

The data points to a diversified series of basic rocks of different ages, developed within the axial zone of the Northern Caucasus, and generally associated with Jurassic volcanism. They undoubtedly contain a fairly widely developed association of high-alkaline Tertiary gabbroids with zeolites.

It follows that the task of classification and petrographic description of vein, intrusive and extrusive rocks of the Main Caucasian axial zone are foremost in the study of Caucasian magmatism.

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Received November 10, 1958

NEAR SURFACE INTRUSIONS AND THE AGE OF THE UYMENSK DEPRESSION GRANITOIDS (GORNYY ALTAY)[1]

by

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This paper deals with the Uymensk depression granitoids belonging to two intrusions of different ages, each of whose component rocks has different geologic, petrographic and, to a certain extent, geochemical properties. The character of the associated ore mineralization of each type is also different. The younger intrusion is represented by near-surface granite porphyry; the older, by coarse-grained biotite and biotite-hornblende granite, components of large massifs.

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A more detailed knowledge of the Uymensk depression magmatism is important not only for the geologic history of the entire Altay-Sayan folded province, but for an evaluation of the mining possibilities of this region.

The Uymensk depression is a well-defined structural zone trending almost from Chuya River (near Chibit village), north-northwest to Biya River, and thence farther northwest, along the middle course of the Kondoma, in Gornaya Shoriya. In the upper course of the Biya, the Uymensk depression joins the Lebedsk depression which becomes its northeastern branch (Fig. 1).

In the northwest, the Uymensk depression borders on Katunsk ridge which originated as an anticlinal structure, as a result of the Salair orogeny. Similar structures border the depression on the northeast. To the east there lies a comparatively little-known region of metamorphic rocks; to the southwest -- an elongated zone of siliceous-carbonate Proterozoic-Cambrian rocks, in a narrow horst. A zone of deep-seated faulting borders the western and eastern margins of the depression.

The stratigraphy of the southern part of the Uymensk depression, built up almost exclusively of Devonian rocks, was described in detail by I. I. Belostotskiy, in 1955.[2] The structure of its northern part has been complicated by the presence of Proterozoic(?), Cambrian and Ordovician rocks. Here lie two large massifs, the Turochak and Sarokokshinsk, chiefly of coarse-grained granite. Similar granite forms only small intrusions in the southern part of the depression.

These features of the northern part of the Uymensk depression are explained by the fact that a Devonian downwarp crosses an anticlinal structural zone here that branches off the Katun ridge to the northeast into Gornaya Shoriya. In the area of this intersection, Cambrian and Ordovician rocks strike northeast, nearly latitudinally, while the flatter Devonian rocks strike north-northwest, nearly meridionally.

Magmatic rocks of the Uymensk depression may be divided into several groups. The oldest of them are Cambrian basic extrusives, developed in its northern part. They have been cut by numerous diabase and porphyry dikes, little known as yet but apparently also Cambrian, at least in part. A sizable part of the depression has been built of Middle Devonian extrusive and sedimentary rocks, with both acid and basic extrusives and tuffs.

Intrusions consist of granitoids, diorite and gabbro, with the gabbro being partly hybrid, a result of assimilation of enclosing rocks by the granite magma, but possibly partly belonging to a different intrusion. The present data do not allow a strict separation of these two groups.

Granitoids, which are much more widely developed here than the basic rocks, make up several intrusions of different ages. The above-mentioned large granitoid massifs are

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[1]Blizpoverkhnostnyye intruzii i vozrast granitoidov uymenskoy depresii (Gornyy Altay).

[2]Belostotskiy's "Uymensk depression" is the southern part of our Uymensk depression.

marked by a fairly complex composition. Besides the dioritic and gabbroid hybrid rocks, developed chiefly along the periphery of these massifs, there are 1) coarse-grained biotite and biotite-hornblende granites, locally porphyritic; and 2) fine- and equigranular to porphyritic granitoids of a different composition and texture. Isolated small granitoid massifs are made up of coarse- to fine-grained or porphyritic varieties.

Coarse-grained granite is widely developed in the northeast of the Gornyy Altay and beyond the Uymensk depression, in plutons which may partly belong to the intrusive complex of the Turochak and Sarakokshinsk massifs. This intrusive complex is termed the Gornaya Shoriya, by geologists of the All-Union Geological Institute. We are still in the dark as to the position of some of these coarse-grained granite massifs, especially those in the metamorphic province and in the south part of the Uymensk depression. There are numerous granitoid pebbles in Ordovician conglomerate along the right bank of the Biya, which suggest a Cambrian or even earlier time of origin of the intrusion. The absolute age of one such granitoid pebble, as determined by the argon method, at the Radio Institute, Academy of Sciences, U.S.S.R., is 395 million years. This strengthens the assumption that ancient granitoids were present in the eastern Gornyy Altay.

Granitoids of the Turochak and Sarakokshinsk massifs are definitely post-Cambrian. They cut and alter Cambrian to Ordovician deposits; there are no pebbles of them in Ordovician conglomerate. There is no concurrence of opinion on the upper age limit of these massifs.

Many geologists correlate them with the Zmeinogorsk or Kalbinsk intrusions of the Gornyy Altay and assign them to the upper Paleozoic. These granites are so designated on the 1:1,000,000 and larger scale maps. According to some other investigators, these intrusions are pre-Devonian, in any event pre-Middle Devonian, and belong to the Caledonian rather than the Variscan orogeny. This last opinion is based on a general analysis of the geologic history of that part of the Altay-Sayan mountain province which includes the eastern Gornyy Altay (east and north of Katun' and Chuya Rivers) and the Gornaya Shoriya. The geosynclinal downwarping of this part of the Altay-Sayan mobile zone began at the close of the Proterozoic or beginning of the Cambrian, with the deposition of the so-called Bartal'sk siliceous carbonate formation. Intensive submarine outpouring of basic extrusives took place at the end of this deposition, which has led to the formation of an extrusive-sedimentary sequence correlative with spilitic keratophyres of other regions. Following the extrusive-sedimentary sequence, terrigenous carbonate deposits were laid down, at the end of the Lower Cambrian, and a Middle Cambrian volcanic-sedimentary sequence of assorted porphyrite, tuff, shale and less commonly limestone.

These formations were strongly deformed by the Salair orogeny when small ultrabasic bodies were probably intruded, as well as diorite and gabbro. All of the sedimentary-extrusive and intrusive formations are characteristic of the initial and early stages of development of a mobile zone, and are marked by a corresponding mineralization.

A substantial change took place as a result of the Salair orogeny, with a sharper differentiation of the area into its structural facies zones. Upper Cambrian, Ordovician and Silurian sedimentary rocks are represented by flysch-like and calcareous shale, typical of middle stages of the mobile zone development. A new folding phase took place during the Ordovician and Silurian, after which a considerable part of the eastern Gornyy Altay and Gornaya Shoriya became a mountain province broken up by depressions which accumulated Devonian sedimentary and sedimentary-extrusive deposits. The Uymensk depression was one of them.

Devonian deposits of the Uymensk depression are represented by vari-colored arenaceous and argillaceous rocks interbedded with products of a chiefly terrestrial volcanism. The Devonian section is somewhat different in the neighboring Ulagan and Chuya-Anuya depressions. Everywhere, however, Devonian rocks have the features of late and terminal stages of the mobile zone development.

The geosynclinal development of the eastern Gornyy Altay and Gornaya Shoriya came to an end in the Upper Devonian. Younger Upper Devonian, Carboniferous and Permian rocks were deposited only in small residual downwarps, west of the meridional part of Teletsk lake and an extension of the Lebedsk depression, in Gornaya Shoriya. These deposits are thin and nearly horizontal.

Thus, the geosynclinal development of the eastern Gornyy Altay, initiated at the close of the Proterozoic, developed along a standard course, and terminated in the Upper Devonian, with transitional deposits intermediate between geosynclinal to platform types.

It may be concluded, then, that the Gornyy Altay granitoids are pre-Devonian, because there are no reasons to assign them to platform-type formations. Field evidence likewise does not favor a post-Devonian age, inasmuch as they do not cut rocks younger than Middle Devonian. However, as indicated

above, there are two age groups of these granitoids -- one represented by coarse-grained, the other by fine-grained and porphyritic varieties. The relationship between the two has been discussed by many authors, without unanimity.

S.A. Yakovlev [10] assigned the granite porphyry to a boundary facies of the Sarakokshinsk massif. This view was shared subsequently by G.M. Saranchina [9], A.A. Menyaylov [8], A.F. Loginov and in part by K.D. Neshumayeva and Zh.D. Nikol'skaya.

A.A. Zenkova and T.F. Vasyutinskaya, who compiled the 1:200,000 geologic map of the central part of the Uymensk depression, assigned the porphyry to one of the last phases of the intrusive cycle which had formed the Sarakokshinsk and Turochak massifs. However, they admitted the possibility that these rocks may belong to a separate and later intrusion.

K.D. Neshumayeva and A.B. Gintsinger, while regarding the Sarakokshinsk and Turochak massifs as a manifestation of a prolonged multiphase intrusive cycle, are inclined to regard the porphyritic rocks as one of its final phases and, in part, its vein derivatives.

In 1953, Ye.P. Zaychenko identified near-surface intrusions within the Turochak granitoid massif, in the northern part of the depression. She termed it a separate granosyenite unit, Middle Devonian in age, and published several papers [4-6] on its petrology and chemistry. However, she had the chance to study only a small area of these rocks confined to the Turochak massif, whose coarse-grained biotite granite she regarded as younger. The work of the field party of the All-Union Geological Institute revealed a much wider development of near-surface intrusions, not only in the area of granite massifs but to the south as well, among Middle Devonian extrusives (Fig. 1). Similar minor intrusions also occur in the northern extension of the Uymensk depression, in Gornaya Shoriya.

Such spatial relationships militate against a close genetic connection between near-surface intrusions and coarse-grained granite, at least against their belonging to a boundary facies of large granitoid massifs.

Direct contacts of the two granitoid types suggest a considerable interval in time between their formation. In a number of localities (Yugola River, 1.5 km above the village of Yugola; upper course of Ashpanak; middle course of Verkhnyaya Ynyrga; middle and upper course of Sarakoksha, etc.), near-surface intrusions are in distinct contact with rocks of the large massifs. In addition, porphyritic rocks cut the vein derivatives of coarse-grained granite (Sarakoksha River, 3.5 km above its confluence with the Ashpanak). In such cases, the adjacent parts of near-surface intrusions often exhibit contact-metamorphic zones, from a few decimeters to 2.5 meters, depending on size of the contacting bodies. These data make it reasonable to believe that the establishment of the major Gornyy Altay massifs was complete, down to the differentiation of vein facies, prior to the origin of near-surface intrusions. They had the time to cool off to the point where they acted as a rigid body for the near-surface intrusions, causing a rapid chilling in the peripheral parts of the intrusions.

Absolute age determinations, obtained under the direction of A.Ya. Krylov in the laboratory of the Radium Institute, Academy of Sciences, U.S.S.R., by the argon method, also indicate a different age for these rocks, with coarse-grained granite being the older, although the difference in figures does not exceed the limit of error. Figures of 328 to 330 million years and 315 million years were obtained for coarse-grained granite and granite porphyry, respectively.

At the same time, petrologic and geochemical features of the large intrusive granitoid bodies are very close to those of near-surface intrusions. The assemblages of similar elements for biotite granite of the Turochak and Sarakokshinsk massifs was determined by spectrum analysis, while silicate analysis has established identical features of their chemistry (supersaturation with alumina and silica, high alkaline content and a low C-feldspathic content). Among the differences in chemical composition is the scarcity of molybdenum, lanthanum and niobium in granite of major intrusions, and a much lower fluorine content (about 0.01 to 0.02%).

Thus, it is not impossible that, granted some age difference, granitoids of both types have originated from a single long-lasting magmatic hearth. This, however, should not have prevented the development of near-surface granitoids into an individual intrusive complex with geologic, petrographic, geochemical and metallogenic features of its own.

Granitoids of near-surface intrusions are represented by areally small bodies of stocks, laccoliths, dikes and dike-like plutons, mostly typical minor intrusions. Dikes occur in isolated bodies or in swarms of 5 to 7 parallel dikes, close together, cutting the extrusive-sedimentary formations and granitoids of larger massifs. There are also conformable, bedded bodies of granitoids in the Middle Devonian extrusive-sedimentary sequence, where they are hardly distinguishable from the extrusives.

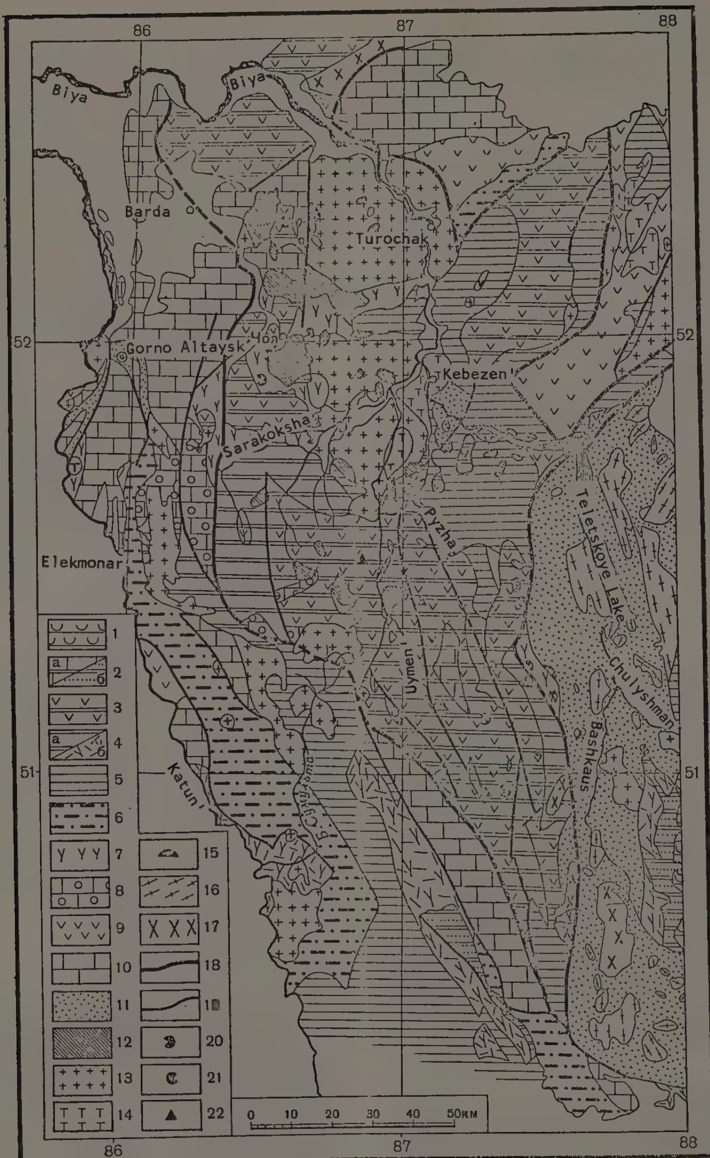

FIGURE 1. Distribution of Middle Devonian intrusions throughout the Uymensk depression.

1 -- variegated arenaceous argillaceous formation (D_2-D_3^1?); 2a -- lagoonal calcareous formation; 2b -- lagoonal arenaceous argillaceous formation (D_2-D_3^1?); 3 -- extrusive-sedimentary red beds (D_2^2); 4a -- red sequence (D_2^1?); 4b -- volcanic formation (D_1-D_2?); 5 -- limestone-shale sequence (O_2-S); 6 -- terrigenous flysh sequence (Cm_3-O_1); 7 -- volcanic-sedimentary sequence (Cm_2); 8 -- terrigenous-carbonate sequence (Cm_1^3); 9 -- porphyritic sequence (Cm_1^2); 10 -- siliceous-carbonate sequence (Pt-Cm_1^1); 11 -- crystalline schist and gneiss; 12 -- granite porphyry, granophyres, granosyenite-porphyry (near-surface intrusions); 13 -- granitoids of the Gornyy Altay; 14 -- quartz diorite, diorite, gabbro; 15 -- ultrabasic rocks; 16 -- granitoids of the metamorphic sequence; 17 -- granitoids of indeterminate age; 18 -- boundaries of the Uymensk depression; 19 -- faults; 20 -- quartz-fluorite mineralization; 21 -- quartz-barite-galena mineralization; 22 -- quartz veins with copper sulfides.

Stocks, laccoliths and dike-like bodies are somewhat elongated laterally, chiefly in a meridional direction. The area of their exposures is usually a few -- rarely more than ten -- square kilometers. The largest massif, southeast of the Turochak intrusion, is about 170 sq km. However, there is no certainty that it is occupied by a single body rather than a chain of smaller massifs separated by the enclosing rocks. This question cannot be answered until some exposures have been cleared. Dikes are 1.5 to 200 m thick, more commonly 10 to 15 m, with a meridional and sub-meridional trend, usually with a steep 65° to 80° SE dip, or sometimes vertical.

The dike form of granitoids, together with the elongation of some of the massifs and the lack of their connection with folded structures in the enclosing rocks suggests their connection with major faults. It is possible that some of the faults which cut these minor intrusions are a rejuvenation of those faults.

Rocks of the near-surface intrusions are fairly resistant to erosion and are the main components of positive relief features. Isolated stock-like bodies are commonly exposed as domes of various size and height. Large massifs or series of stock-like bodies are exposed in high swells. Granitoid dikes commonly stand out as crests, from tens to a few hundred meters long; they are especially conspicuous in biotite granite which is readily weathered.

Contact metamorphism between near-surface intrusions and the emplacing rocks occurs in a narrow zone, several decimeters to a few meters wide, as a recrystallization of contact extrusive-sedimentary rocks, with the formation of hornstone and occasional scarn and epidotization.

Petrographically, granitoids of small intrusions are an isolated facies group with well-developed features indicative of their near-surface origin. Most of them are made up of monotonous brick-red, pinkish-gray and yellow porphyritic rocks.

S.A. Yakovlev, A.A. Zenkova and some other investigators regard the typical representatives of these rocks as spherolitic porphyry, quartz albitophyre, other porphyries, etc. G.D. Afanas'yev [1] noted the "granite porphyry aspect" of the Sinyukha Mountain granitoids of this type and the presence of "typical micropegmatites" among them. Inasmuch as the representatives of these rocks are of the same intrusive origin and of the same, or similar, chemical composition, and differ only in the texture of their matrix and the amount and size of phenocrysts, it is expedient to call them all granite porphyry, after A.N. Zavaritskiy [3], using the texture of the matrix as a criterion. Thus, granite porphyry with a pseudospherulitic matrix should be called pseudospherulitic granite porphyry, etc.

The most common varieties of small intrusive granitoids are micropegmatitic and pseudospherulitic granite porphyries, with less common felsitic granite porphyries, granophyres and grano-syenite porphyries.

Macroscopically, they are dense rocks, chiefly of a porphyritic texture, with phenocrysts generally rectangular to square, from fractions of a centimeter to 1.5 cm large. It is not uncommon that several such phenocrysts grow together in a glomerophyric texture. In composition, they are plagioclase, K-feldspar and quartz, with dark minerals comparatively rare, being represented generally by biotite.

A zonal distribution of rocks is characteristic of several massifs. The central parts are made up of better-crystallized varieties, with up to 40% porphyry phenocrysts, 0.5 to 0.6 cm on the average. Under the microscope, they exhibit the best-crystallized micropegmatitic and micrographic structure of the matrix. Miarolitic cavities are fairly common in these segments of massifs. They are 0.7 to 0.8 cm in diameter, with walls covered with a bloom of brown iron oxide and opaline quartz. Central parts of large massifs also consist of fine-grained granite.

The amount and size of phenocrysts decrease toward the periphery of the massifs. In the near-contact zone, they are 0.1 to 0.3 cm large and their amount drops down to 10 or 12%; locally, they are altogether lacking. Pseudospherulitic and felsitic textures are especially characteristic of this portion of the massifs.

Zonation is less distinct in dikes; instead, there is a contact-metamorphic zone, 20 to 30 cm wide in small dikes, to a few meters in larger ones. Rocks of this zone often lack the porphyry phenocrysts; they have a felsitic matrix and streamline, banded texture.

A number of thin sections, taken across the strike of small intrusions, reveal regular changes in the matrix. A cryptocrystalline felsitic matrix of a quartz-feldspar composition predominates in the peripheral zone. Nearer the center, pseudospherulites of the same quartz-feldspar composition appear in the felsitic groundmass. They are isolated, with the intervals between them filled with the felsitic mass carrying fine scales of biotite, sericite, chlorite and occasional epidote. Radially-fibrous pseudospherulites of joined threads of quartz and albite predominate.

Farther away from the contact, dendritic pseudospherulites appear alongside the fibrous ones. They are formed by micropegmatitic growths of quartz and feldspar, with interstitial micropegmatitic and micrographic structures appearing in the felsitic matrix. In more remote segments, pseudospherulites lose their definite outline and gradually disappear to give place to micropegmatitic structures. In thin dikes, such changes are less gradual and not as definite.

Combinations of different structures within the same body are not uncommon alongside the regular change in the distribution of structural varieties.

The porphyritic texture of rocks and the rapid change in the texture of the matrix suggest extremely inconsistent physical and chemical conditions of crystallization for these rocks, a characteristic of hypabyssal salic, near-surface rocks.

Microscopic study shows that rocks of minor intrusions have much in common with Upper Devonian extrusives. Porphyritic phenocrysts of various granite porphyries and granophyres contain orthoclase, acid plagioclase, quartz and less commonly biotite. Orthoclase and albite (usually secondary) clearly predominate in the syenite porphyry phenocrysts, with quartz lacking and biotite as rare as it is in granite porphyries. Quantitative ratios of these minerals are variable, often within a single body. However, orthoclase usually is the predominant phenocryst mineral.

Orthoclase occurs in isometric crystals with corroded edges, commonly overgrown with quartz, which creates the effect of a fringe about them. The orthoclase is strongly pelitized, replaced locally by checkerboard albite and commonly carrying perthitic growths of albite.

Plagioclase is much less common in porphyritic phenocrysts, being usually represented by an acid variety corresponding to albite, with a 5 to 10% content of anorthite. More basic varieties are much less common. Plagioclase is commonly strongly sericitized.

Quartz phenocrysts are often rounded or polygonal, with fused edges containing inlets of groundmass. In places it has wavy extinction.

Biotite is present in elongated scales, almost everywhere strongly chloritized. The unaltered varieties are dark-colored, strongly pleochroic, from red-brown along γ to colorless along α. Biotite inclusions are commonly associated with ore and some accessory minerals (zircon, apatite).

Acid extrusives of the Uymensk depression, corresponding in composition to quartz-free porphyry, albitophyre and quartz porphyry, carry similar porphyritic phenocrysts with very similar to identical properties.

As already noted, the matrix structure of porphyries is very inconsistent, with micropegmatitic and pseudospherulitic varieties the most common.

The micropegmatitic texture usually is accompanied by the micrographic, both being formed by the regular intergrowth of quartz in orthoclase. In different parts of the rock, quartz forms worm-like, rounded to irregular growths (micropegmatitic texture) or else geometrically regular trapezoidal and rectangular forms (micrographic texture). The quartz growths range in size from thousandths to tenths of a millimeter, depending on the degree of crystallization of the rock. This texture is peculiar to those parts of an intrusion farthest away from the contact.

The pseudospherulitic texture, too, is common, mostly in conjunction with the others. Fibrous, radial, and dendritic forms have been recognized among pseudospherulites. The first consists of fine straight fibers of essentially K-feldspar and quartz, usually with a cross-like extinction. Dendritic varieties are represented by branching growths of quartz and feldspar, with the size and amount of the former increasing toward the pseudospherulite periphery. They are characterized by wavy or sectorial extinction.

"Embryos," or crystallization centers of fine grains of quartz, ore mineral, or feldspar, are nearly always discernible in pseudospherulites. The diameter of pseudospherulites usually is 0.2 to 0.8 mm, rarely one mm. The space between them is filled with a felsitic or micropegmatitic substance of a quartz-feldspar composition. Pseudospherulites account for 55 to 60% of the matrix, with much less in some varieties. Interstices between them are filled with scales of muscovite, chlorite, albite, and epidote grains, measured in hundredths of a millimeter.

The pseudospherulitic structure is also characteristic of the extrusive representatives of Devonian magmatism. Here, however, it is somewhat different from similar textures of subvolcanic rocks. The extrusive varieties are marked by a predominance of less individualized pseudospherulites of a quartz-feldspar substance with a refractive index below 1.54, so that minerals cannot be identified even at high magnification. Metaspherulite textures are not uncommon, here.

The felsitic texture predominates in

Table 1

Chemical composition of Middle Devonian magmatic rocks.

Oxides	1 Granophyric gran. por. Ugul River	2 Granophyre, Mt. Tsygan	3 Ps. spheru. granite por., Sartakol R.	4 Granophyre, Mt. Kyzyl-Tash	5 Pseudospherulitic gran. por. Sarakoksha River	6 Ps. spheru. granite por., Baushka vil.	7 Granophyric granite por., Sarakoksha R.	8 Micropegmatitic gr. por. Seyka River	9 Micropegmatitic gran. por. Tazhervakh-Zapegala R.	10 Pseudospherulitic granite porphyry
SiO_2	75,33	74,96	75,24	77,08	74,64	73,50	75,69	72,49	73,21	69,85
TiO_2	0,13	0,10	0,10	0,10	0,18	0,13	0,17	0,26	0,22	0,52
Al_2O_3	12,78	11,47	13,58	11,14	13,42	15,09	12,09	13,99	14,35	14,29
Fe_2O_3	0,74	1,87	0,65	1,76	0,99	0,99	2,55	0,67	1,64	1,04
FeO	1,48	1,20	0,56	1,26	0,57	0,48	0,52	1,88	0,67	2,18
MnO	0,03	—	0,04	Traces	0,06	0,06	0,18	0,05	0,05	0,06
P_2O_5	—	—	—	Traces	—	0,04	0,03	0,04	—	0,07
CaO	0,69	0,75	0,64	0,98	0,84	1,74	0,20	1,10	0,61	1,55
MgO	0,20	0,17	0,33	0,24	0,29	0,45	0,18	0,40	0,20	0,70
Na_2O	3,64	2,43	4,42	2,38	3,41	4,42	3,76	4,26	4,06	4,06
K_2O	4,40	5,21	3,06	4,68	4,07	2,75	4,33	3,57	4,58	4,16
H_2O	0,20	1,20	0,22	1,10	0,17	0,17	0,21	0,14	0,08	0,20
Loss in heating	0,53	0,56	1,01	0,50	1,57	0,43	0,25	0,98	0,64	1,28
Sum	100,15	99,92	99,85	100,22	100,21	100,25	100,16	99,83	100,31	99,96

Oxides	11 Felsitic granite por. Sarakoksha R.	12 Micropegmatite granite porphyry	13 Micropegmatite gran. por. Argundu River.	14 Granite por., Salop range	15 Granosyenite por., Arkhipova flood pl.	16 Granosyenite porphyry, Ashpanak R.	17 Granosyenite porphyry, Mt. Tsygan	18 Quartz syenite porphyry, Chistinka settlement	19 Keratophyre, Ayya massif	20 Keratophyre, Turochak massif	21 Albitophyre, Ayya massif
SiO_2	75,67	70,40	71,04	74,56	71,38	73,58	72,06	70,68	76,52	68,64	73,80
TiO_2	0,15	0,35	0,36	0,07	0,33	0,23	0,26	0,09	0,09	0,14	0,04
Al_2O_3	11,45	13,45	14,22	14,37	14,38	14,71	14,88	14,63	11,28	14,57	13,53
Fe_2O_3	2,39	1,54	0,89	—	0,97	0,97	1,30	1,06	0,60	1,76	0,71
FeO	1,11	2,18	1,56	1,54	2,61	1,30	1,45	3,17	1,20	3,17	3,81
MnO	0,04	0,07	0,05	—	0,08	0,03	0,06	—	—	—	—
P_2O_5	0,02	0,07	—	—	0,07	0,18	0,04	Traces	Traces	Traces	—
CaO	0,40	1,35	0,84	1,11	1,15	0,33	0,22	1,11	1,02	1,97	1,29
MgO	0,25	0,60	0,47	0,59	0,83	0,29	0,39	0,87	0,55	0,66	0,71
Na_2O	3,95	4,09	3,20	3,87	3,75	3,82	4,00	3,80	3,18	4,59	4,13
K_2O	3,12	4,47	6,10	3,55	3,88	3,39	4,02	3,67	4,63	4,24	2,08
H_2O	0,49	0,14	0,38	0,24	0,23	0,19	0,38	0,24	0,54	0,06	0,12
Loss in heating	0,98	1,29	0,79	0,40	0,52	1,02	1,02	0,74	0,66	0,44	0,12
Sum	100,92	100,00	99,90	100,30	100,18	100,04	100,08	100,06	100,27	100,23	100,34

Analysed in the chemical laboratory of the All-Union Geological Research Institute: 2, 3, 5, 6 -- by V. M. Kovyazina; 1, 9, 13 -- by O. G. Arbidan; 7, 8, 10, 13 -- by V. A. Yusova; 15, 16, 17 -- by M. T. Selyugina; analyses 14, 18, 19, 20 have been borrowed from E. P. Zaychenko's paper [4].

Table 1, continued

Numerical characteristics, after A. N. Zavaritskiy

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
a	13,7	12,5	13,5	11,5	12,9	13,1	13,9	14,2	15,0	14,9	12,6	15,2	15,5	12,9	13,2	12,5	13,9	13,0	13,2	15,9	11,3
c	0,8	0,9	0,8	1,0	0,9	2,0	0,3	1,3	0,7	1,9	0,5	1,2	0,9	1,2	1,3	0,4	0,3	1,3	0,6	1,6	1,5
b	3,3	3,7	3,8	4,5	4,3	4,3	3,9	4,3	4,2	4,3	4,8	4,7	4,7	5,3	7,2	7,7	7,6	8,3	3,2	4,6	7,4
s	82,2	82,9	81,9	83,0	81,9	80,6	81,9	80,2	80,1	78,9	82,1	78,9	78,9	80,6	78,3	79,4	78,2	77,4	83,0	77,9	79,8
f'	58,0	74,5	27,6	75,5	32,9	30,8	65,0	53,9	46,0	67,7	62,2	70,4	56,0	26,8	44,5	25,9	31,6	44,4	53,2	72,5	50,0
m'	10,0	7,3	13,8	9,4	10,9	16,9	8,3	15,3	8,0	26,2	8,1	21,1	17,0	14,6	19,1	5,8	8,6	17,5	30,0	16,5	13,69
a'	32,0	18,2	58,6	15,1	56,2	52,3	26,7	30,8	46,0	6,1	29,7	—	28,0	58,6	36,1	68,3	59,8	38,1	—	—	36,1
c'	—	—	—	—	—	—	—	—	—	—	—	8,5	—	—	—	—	—	—	17,0	11,0	—
n	55,0	41,5	68,2	43,8	56,1	71,0	57,0	64,5	57,0	59,9	65,9	57,9	44,0	61,6	59,4	62,9	60,7	61,1	51,5	62,2	75,8
φ	16,0	43,6	13,8	41,5	18,7	18,5	53,3	12,3	31,1	18,4	40,5	25,3	15,0	—	11,0	[illegible]0,0	13,6	9,5	17,0	24,2	6,5
t	0,03	0,1	0,1	0,1	0,2	0,2	0,2	0,3	0,2	0,5	0,1	0,4	0,4	—	0,3	0,2	0,3	0,1	0	0,1	0
Q	36,2	39,9	36,0	42,1	37,1	33,0	35,7	30,7	29,5	26,1	38,5	26,2	25,9	[illegible]4,2	28,9	33,4	28,3	27,1	39,0	22,4	35,5
$\frac{a}{c}$	17,1	12,5	17	11,5	12,0	6,5	46,3	10,9	21,4	8,2	25,2	15	17,2	10	10,1	31,2	46,[illegible]	10	22,0	9,9	7,5

Note: Comma represents decimal point.

peripheral parts of areally small intrusions, where it is represented by a poorly defined substance having a quartz-feldspar composition. Locally, it has been better crystallized, with discernible fine (0.01 to 0.02 mm) inclusions of quartz, K-feldspar, plagioclase and with ever-present fine, strongly elongated dark brown-red biotite and pulverized ore minerals. The felsitic texture occurs most commonly in conjunction with others. It is also present in some extrusive rocks.

The groundmass of granosyenite porphyry is somewhat different, being formed by fine (0.08 to 0.1 mm) short prismatic crystals of albite (Nos. 4 to 6), amounting to 65 or 70% of a micropegmatite matrix. The quartz and K-feldspar intergrowth is measured in thousandths of a millimeter. Micropegmatite growth of quartz and orthoclase usually forms a halo about albite crystals.

Secondary minerals of subvolcanic intrusions are chlorite, sericite and occasionally hydrobiotite.

A study of powder of various rocks from smaller intrusions has indicated a fairly consistent composition of accessory minerals, each one amazingly persistent in its habit, color and other properties, regardless of petrographic associations. This is especially true for zircon, fluorite, barite and apatite.

Zircon is one of the most common accessory minerals. It occurs in fine dipyramidal prismatic crystals (0.3 x 0.075 mm), some crystals markedly elongated along the third crystallographic axis. The crystal aspect is determined by a wide development of prism faces (110), with second order prism faces (100) always present, although not as well developed, and facets of an obtuse dipyramid (111). Zircon is light yellow, of various intensity, locally orange, semi-transparent, sometimes strongly fractured, with uneven crystal faces, and a vitreous luster. It commonly has inclusions of dark to brown minerals.

As seen in thin sections, zircon is frequently associated with ore minerals and biotite; it is not as common in quartz-feldspar matrices.

Fluorite is common in small intrusions, but usually in smaller amounts than zircon. It occurs mostly in isometric grains, up to 1 mm in length, without definite crystallographic faces. Its pale violet coloring is spotty, but there are less common colorless varieties faintly-tinted yellowish. Spectrum analysis shows the presence of lead, molybdenum, copper, iron, titanium, manganese, zircon, yttrium and strontium.

Barite is common, but in small amounts, occurring in amorphous grains, up to 0.8 mm in diameter.

Apatite is represented by hexagonal prismatic crystals. It is colorless, semi-transparent with corroded crystal faces. Microscopic study reveals spatial association of apatite grains with zircon and ore minerals, although it may occur in a granite porphyry matrix, as well. It is common in porphyries of small intrusions.

Octahedrite (anatase) is a widely distributed accessory mineral of small intrusions. It has been observed in all specimens, but only in isolated fragments of prismatic crystals, a few tenths of a millimeter in diameter. It is dark green, usually non-transparent.

Tourmaline forms rare crystals and does not occur in all samples. It is black, with well-developed striation of faces; up to 0.5 mm in diameter.

Sphene is common. Its content is very variable, from a few single grains in some samples to a few percent of the heavy fraction in others. It occurs mostly in fragments of tabular crystals. It is dark in color, at times with a gray leucoxene bloom on the grains.

Pyrite is a fairly common accessory mineral in dikes. It forms cubic crystals with well-defined striation; average diameter, 0.5 mm, and up to 2 mm in isolated crystals.

Magnetite is always present, and in fairly large amounts, mostly in well-formed octahedral crystals, 0.2 to 0.5 mm.

Ilmenite is less common, occurring in black tabular crystals.

Isolated grains of galena have been encountered in many granitoid specimens of small intrusions where it forms fine crystals, 0.2 to 0.3 mm.

Native lead has been found in two samples (one grain in each), represented by fine hook-like inclusions.

Garnet, scheelite, orthite and monazite have been observed in single grains, in a few samples.

Granitoids of the various massifs under study, and acid extrusives of the Uymensk depression, are very similar in their chemical composition, as it appears from the chemical analyses (Table 1) and the geochemical diagram constructed by the Zavaritskiy method (Fig. 2). Most of the analysed rocks occupy an intermediate position between the normal (calcium-sodium) and alkaline series, approaching the latter. The projections

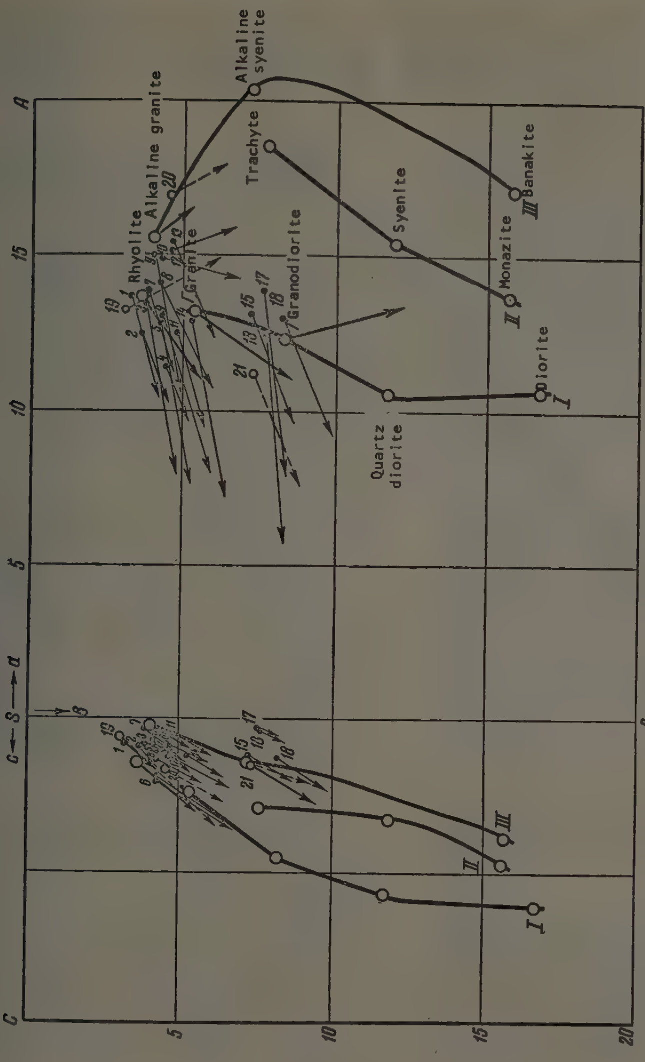

FIGURE 2. Geochemical diagram of hypabyssal rocks of the Uymensk depression.

Average composition (after R. Daly): I -- normal series; II -- syenite-monzonite series; III -- alkaline series.

of figurative points for hypabyssal and extrusive rocks are concentrated near each other. In plane ASB, points of the composition of magmatic rocks of the Uymensk depression form two assemblages, corresponding to granite porphyry and granosyenite-porphyry.

Extrusive rocks display the same tendency. Unfortunately, there are only three analyses of them, two of which (19 and 20 in Fig. 2) fall in the granite porphyry assemblage, while one (21, Fig. 2) falls close to the granosyenite porphyry assemblage.

The common and well-defined chemical features of these rocks are as follows:

1. Strong supersaturation with alumina. In most cases, the value of coefficient a is not less than 25, in some cases reaching 68. On the diagram, this feature is expressed by the vector direction toward axis SB, in plane ASB.

2. Abundance of silica whose content varies but slightly. Most common are granite porphyry with 73 to 74% silica, and granosyenite porphyry with 70 to 73%.

In Zavaritskiy's classification, all analyses fall into a class of rocks supersaturated with silica; therefore their points are concentrated near the top of the diagram (Fig. 2). Granosyenite porphyries containing somewhat less SiO_2 are farther away from top S.

3. A higher alkalinity. The sum of alkalis in most analyses is 7.5 to 9%. Ratio a/c, for granite porphyry, is never less than 8, reaching 30 for granosyenite porphyry. In Zavaritskiy's classification, they correspond to high alkali rocks (a/c > 7). The higher alkalinity of rocks is fairly well reflected on the diagram, with all points corresponding to the chemical analyses of granite porphyry located in plane CSB, to the right of the normal series tier and concentrated about the average composition point for alkaline granite, according to R. Daly. The figurative points for granosyenite-porphyry lie to the right of those for alkaline rocks, and near the alkaline syenite point of Daly.

Varieties with potassium somewhat predominating among alkalis are widely distributed among granite porphyries. Granosyenite porphyries are marked by a higher content of potassium and sodium oxide, with the latter slightly predominating. This is reflected on the diagram by steeper slopes of the granosyenite porphyry vectors as compared with the granite porphyry (plane CSB).

4. The low feldspathic lime content. Factor c does not exceed 0.5 to 0.8%, in most analyses; accordingly, points of plane CSB are crowded near axis SB.

5. The high fluorine content in samples, 0.06 to 0.08% on the average.

The chemical features of acid extrusives and near-surface intrusions of the Uymensk depression were spectrum analysed, by the semiquantitative method, in the laboratory of the All-Union Geological Research Institute (VSEGEI). It has been found that both intrusive and extrusive rock groups contain identical assemblages of elements such as Mn, Ti, Fe, Ga, Sr, V, Zr, Nb, Be, La and especially Cu, Pb, Mo and Ba. The content of the last four elements is several times higher than the average for acid rocks, according to A.P. Vinogradov [2]. For example, the molybdenum content in the acid representatives of extrusive rocks and in near-surface intrusions varies from 0.001 to 0.003%, while the average for this element, according to A.P. Vinogradov, is 0.00019%.

Spectrum analysis of individual minerals (magnetite, zircon, sphene, apatite and fluorite) from granite-syenite and granosyenite-porphyry, separated from powder with an electromagnet and heavy liquids (with a microscopic study of material so selected), has revealed the presence of copper; lead, molybdenum and barium, in all samples. Zircons had the highest lead content (up to 1.0%) and molybdenum (0.1 to 0.3%).

All these data on the geology, petrography and geochemistry of near-surface intrusions make it possible to distinguish them from other intrusions of eastern Gornyy Altay, and particularly from the Gornyy Altay granite. The same data point to a similarity of smaller intrusions to Upper Devonian extrusives. The association of these intrusions with the area of distribution of Middle Devonian variegated extrusives is well illustrated in Figure 1. Similar granitoids have been described by Yu.A. Kuznetsov [7] from another region of the Gornyy Altay, where they are likewise associated with an area of development of Middle Devonian extrusives.

All this suggests a genetic connection between the granitoids in question and Middle Devonian extrusives, and a subvolcanic origin for at least some of the granitoid massifs.

Ore deposits, associated elsewhere with such intrusions, have not been found in eastern Gornyy Altay, as yet. These ore deposits are chiefly sulfide-fluorite, galena-barite and copper-molybdenum mineralizations.

Sulfide-fluorite ore exposures are widely developed throughout the Uymensk depression. As a rule, they are too small to be of commercial interest.

Galena-barite ore exposures are less common, occurring chiefly in the central part of the depression, in two localities where they are associated with the outer contact zone of felsitic granite porphyry. The mineralization takes place in veins, chiefly of barite and quartz carrying incrustations and druses of galena and less commonly of sphalerite, chalcopyrite and pyrite. Occasionally, a very small amount of fluorite is present in incrustations. The near ore rocks have undergone intensive silicification and sericitization. Barite veins, barren of ores, are much more common.

Copper-molybdenum mineralization is represented by a small ore exposure in the central part of the depression, at the southwestern outer contact of the pre-Devonian Sarakokshinsk massif, in the lower course of the Kul'bich. The enclosing rocks have been strongly altered: i.e., silicified and sericitized.

Besides these instances of mineralization, inconspicuous skarning of enclosing rocks with the formation of garnet, vesuvianite and epidote has been observed in several outer

contacts of near-surface intrusions. These rocks are associated with thin veins, fine incrustations and small druses of hematite. Other ore minerals include small amounts of galena, chalcopyrite, pyrite, and molybdenum as revealed by spectrum analysis.

Extrusive facies of Devonian magmatism have a poor development of sulfide mineralization in incrustations.

All these types of mineralization are closely related geochemically. A group similarity in mineral composition has been observed in the sulfide-fluorite, galena-barite, and copper-molybdenum types, with barite, fluorite, galena, chalcopyrite and molybdenite present in all of them. Their main difference is in the quantitative ratio of these minerals. The hematite mineralization, too, is accompanied by a small amount of galena and chalcopyrite incrustations, with the presence of molybdenum detected by spectrum analysis.

The petrographic and geochemical study of Middle Devonian magmatic rocks of the Uymensk depression points to a genetic connection between the above-named mineralization types and near-surface intrusions.

The principal minerals of hydrothermal formations -- barite, fluorite, galena, and chalcopyrite -- are widely developed among the accessory minerals of near-surface intrusions. In addition, their contamination with lead, molybdenum, barium, and fluorine has been revealed by chemical and spectrum analyses. The above-mentioned close spatial relationship of mineralization with near-surface intrusions is well demonstrated on the geologic map of the Uymensk depression (Fig. 1).

Minor intrusions along fractures, related to surface volcanism, are typical of the later stages of development of many mobile zones. This is the position of the near-surface granitoids in the geologic history of eastern Gornyy Altay. The related phenomena of endogenic mineralization, probably not fully understood as yet, also belong to typical development of the later stages.

The relationship of near-surface intrusions with extrusives suggests nearly the same age for both, i.e., the near-surface, partly subvolcanic, intrusives are most likely post-Devonian. This is in accordance with the absolute age determination data.

As to the upper age limit for the Turochak and Sarakokshinsk coarse-grained granite, it appears to be pre-Middle Devonian, because Middle Devonian near-surface intrusions are younger than the Gornyy Altay complex. As pointed out before, the geochemical similarity of granite porphyry and coarse-grained granite suggests their co-magmatism. Consequently, an early Devonian age for the large granitoid massif, prior to the beginning of Middle Devonian, should not be ruled out. On the other hand, the fact that the Turochak massif is enclosed in Cambrian and Ordovician sedimentary-extrusive and sedimentary rocks suggests its formation during the period of main folding, i.e., pre-Devonian. In the character of the associated mineralization -- represented among other types by tungsten ore in greisens -- the Turochak and Sarakokshinsk massifs are most reminiscent of the middle stages of development of mobile zones. If this is true, the intrusion took place during the Ordovician or Silurian.

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Received October 22, 1957

DUNITES OF THE BORUS RANGE AND THEIR ORIGIN[1]

by

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Dunites of the Borus Range (western Sayan) are described. On the basis of this study, the author believes in a metasomatic origin of dunite and associated enstatite. The two rocks are regarded as products of two stages of the same metasomatic process.[2]

* * * * *

THE ORIGIN OF DUNITE

The problem of the origin of dunite has never been regarded as definitely solved, although both dunites and associated mineralization have always been regarded as magmatogenic.

This opinion was shared by most students who differed only on details, e.g., whether there is a dunite magma proper. If not, what are the processes responsible for the differentiation of dunite from a magma of another composition, and what is the place of dunite in the crystallization sequence of rocks, such as the components of structurally complex massifs of the Uralian gabbro-peridotite type?

The existence of a dunite magma proper was supported by only a few workers. In recognizing an olivine magma, Dobré believed that it underlay the rocks of the crust. In considering that dunite commonly forms fairly large independent bodies. F. Yu. Levinson-Lessing was inclined at one time to recognize the existence of dunite magma [13]. In the thirties and forties, A.N. Aleshkov [1, 2] insisted on the existence of a dunite magma, believing it to be the only possible one.

The vast majority of other authors held to a different opinion. N. Bowen, P. Niggli, J. Vogt, M. Schweig, P. Eskola, D. Spurr, S.R. Nockolds, F.Yu. Levinson-Lessing, A.N. Zavaritskiy, G.L. Padalka, P.I. Lebedev, Ye.A. Kuznetsov and others believed that dunites were products of fractional crystallization of a basalt magma. The proponents of a heterogenous origin of ultrabasic magmas (A. Holmes, G.G. Hess, Yu. A. Kuznetsov, M.A. Kashkay, G.V. Pinus and others) believe it possible to relate the origin of dunite not only to a basalt magma but to a special ultrabasic one, as well.

It is well known that the partisans of the fractional crystallization hypothesis, too, differ among themselves. Serious difficulties did arise in the explanation of the origin of dunite dikes. While the majority believed that dunites were first to be formed in the complex system of the Uralian type gabbro-peridotites, where they form the core of "confocal" massifs, L. Duparc argued that dunites had crystallized last. Only in this way, he held, could a dunite dike have originated. He was opposed by F. Yu. Levinson-Lessing who pointed out that the position of the rocks in the massifs contradicted that view.

The most ardent partisans of the gravitational hypothesis -- M. Schweig (following F.Yu. Levinson-Lessing, [13]), G.L. Padalka [16] and others -- attached great significance to the settling and remelting of heavy olivine crystals in the depths of a magmatic hearth. Such remelting, of course, called for very high temperatures, the presence of which was categorically denied by the main proponent of this hypothesis, N. Bowen.

Further study of ultrabasic rocks, as was to be expected, introduced new petrologic data and exposed the inconsistence of the gravitational hypothesis. Many leading authorities (A.N. Zavaritskiy, A.P. Karpinskiy, P. Eskola, D. Spurr, etc.), when confronted with new evidence, renounced this hypothesis as an explanation of the origin of dunite. A.N. Zavaritskiy was forced to admit the existence of a "true dunite magma." This

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[1] Dunity khrebta borus i ikh proiskhozhdeniye.

[2] The author has interesting data on the origin of dunite and associated rocks. However, his conclusions are open to argument and require further refinement, supported by broader evidence. Russian Editor.

opinion was concurred with, later on, by D.S. Belyankin [3], who went into more detail. A substantiated criticism of the gravitational hypothesis was provided by K.N. Fenner and V.N. Lodochnikov [12].

An explanation of the origin of dunites from an ultrabasic magma, by means of the fractional crystallization and gravitational hypothesis is just as impossible. This, apparently an obvious fact, escapes some of the modern petrologists. For instance, G.V. Pinus and I.V. Volokhov [15], having studied ultrabasics of Tuva, categorically rejected the hypothesis of a gravitational differentiation of basalt magma. At the same time, they invoked that hypothesis in their explanation of the origin of dunite and pyroxene, within the ultrabasic zones, without proving its applicability to ultrabasic magma.

Thus, these authors, having correctly established one of the patterns in the rock distribution within ultrabasic belts -- but limiting themselves to a magmatic concept of their origin -- are unable to explain the mechanism of those processes responsible for an ultrabasic magma yielding pyroxenite, harzburgite, and dunite.

During the second half of this century a number of papers were published advancing a metamorphic origin for certain dunites. In 1952, N.M. Uspenskiy [21], after a study of confocal ultrabasic massifs of the Urals, concluded that dunites and other rocks of those massifs had originated metasomatically. Broken ultrabasics of the Baranchinskiy massif, described earlier by A.N. Zavaritskiy as "conglomeratic serpentine," and as "eruptive breccia" by O.A. Vorob'yeva, were regarded by S.V. Moskalev [14], in 1954, as typical replacement breccia where pyroxenite had been metasomatically replaced by dunite. In 1956, A.A. Kadenskiy [7] described as pseudodunite the olivine rocks of Malka River, which he believed to have originated as a result of redistribution of MgO and SiO_2 in an ultrabasic body, under the action of superheated solutions. H. Bowen and O. Tuttle [4], after a detailed study of system MgO-SiO_2-H_2O, demonstrated that serpentine, enstatite and olivine are easily synthesized hydrothermally, and that there could not be any silicate melt below 1,000°C. Their main conclusion is as follows: "Field observations, which suggest a low temperature for the process, as well as our own experimental data, rule out the possibility of a liquid-magmatic state."

Thus, the long road in search for a solution to this problem has led petrologists to the most contradictory views.

DUNITES OF THE BORUS RANGE

1. Position and Structure of Dunite Bodies

These dunites are located within an ultrabasic massif, chiefly of serpentinized peridiorites and and serpentinites. The massif is a large dike-like body, trending N - 40-45° - E, conformably with ancient, strongly disturbed rocks. To the northwest, the ultrabasics are in contact with green-gray to gray phyllitized shale and carbonaceous shale, interbedded with quartzite. This sequence is supposed to be Lower Cambrian (the Chinginsk formation, according to A.G. Sivov).

Southeast of the dunite bodies, there is a small area of ultrabasics in contact with peculiar albite-epidote-chlorite gneiss. Farther away, in both directions along the strike, their ablite porphyroblasts disappear, and they look like green crystalline schist, similar to those of the Dzhebashsk formation [19].

The dunites make up two comparatively small bodies, with an overall area of about 5 sq km. One of them lies wholly among serpentinized peridotite and serpentinite, while the other is associated with the latter's contact with albite-epidote-chlorite gneiss. They are separated by serpentine rock, 250 to 300 m thick.

The relationship between Dunite Body One and the serpentinized peridotite is very complex. It is, therefore, difficult to identify the contact; there is rather a contact zone, 50 to 100 m thick, marked by a wide development of dike-like bodies of one of these rocks in the other. The number of dunite lenses increases toward the center of the dunite body, while the number and size of serpentinite lenses decreases. The thickness of dunite lenses also increases with depth.

Dunite Body One is characterized by a parallel system of vein-like pyroxenite bands, generally 2 to 5 cm thick, locally 30 to 40 cm. With vertical dips, the banding has a nearly meridional strike (N-10°-E to N-10°-W); this is also the strike of dunite lenses in serpentinite. Thus, the textural elements of Dunite Body One are transverse to the strike of the ultrabasic massif as a whole (Fig. 1).

Dunite Body Two is rounded in plan, forming a small dome-like elevation. Its relationship with the enclosing rocks is not clear, because of a blanket of rock slide covering it. It appears that the convex southeastern side of the dunite body enters the albite-chlorite-epidote gneiss. A distinguishing feature of this type of rocks is a wide development of ptygmatic textures brought about by the presence of folded pyroxenite veins

FIGURE 1. Generalized geologic map of the Borus Range.

1 -- dunite; 2 -- albitite; 3 -- talc-carbonate rocks; 4 -- granite of the Dzhoyskaya intrusion type; 5 -- serpentine rocks and serpentinized peridotite; 6 -- phyllitized shale and carbonaceous shale interbedded with quartzite and basic extrusives (Chinginsk formation); 7 -- crystalline schist of the Dzhebashsk formation; 8 -- albite-epidote-chlorite gneiss.

in thoroughly crystallized dunite.

The vein-like bands and folded veins of pyroxenite are associated only with these dunite bodies and have not been found elsewhere among serpentinized ultrabasics. This fact, along with the others, suggests a close genetic association of pyroxenite and dunite.

2. Petrographic Description

a) Dunites and related varieties.

Besides olivine, dunite nearly always carries rhombic pyroxene and impurities of a chrome-spinel rock, talc, and occasional serpentine. Not a single thin section contained monoclinic pyroxene. The rhombic pyroxene content, however, is subject to considerable fluctuation (from 0 to 35%), which is the basis for the differentiation between dunite, enstatite and harzburgite.

Microscopically, all these rocks have a fresh aspect; their texture is medium- to coarse-crystalline, equigranular, massive, rarely coarsely schistose. Their color does not depend on the pyroxene content; it is usually greenish yellow to grayish yellow, with gradual changes to dark gray and black with a faint greenish tint. Less common are yellow and pale green varieties in which the olivine has a vitreous luster. Besides the olivine, there are fine incrustations of a chrome-spinel rock whose black grains stand out especially well against the drab-yellow weathered surface. Chrome-spinelids are fairly evenly distributed, locally in chains and bands, but rarely in massive segregations. No large accumulations of them have been found in the main dunite outcrop.

Pyroxenes are inconspicuous on a fresh surface, but stand out well on a weathered surface, being more resistant than olivine.

The microscope usually reveals the uneven grain of dunite. Alongside large olivine grains, 5 to 6 mm in diameter and attaining 14 mm, there are nearly always small grains present, tenths and hundredths of a millimeter in length. The large olivine grains are usually linked into an intricate broken line. Protuberances and apophyses of some of the grains are commonly deeply imbedded into others, locally splitting them. The bulk of fine grains are segregated between larger grains. Their form is usually irregular, often wedge-like or in short veinlets. Such a relationship between the grains of the same mineral is always suggestive of the existence of several generations.

Large olivine grains in some thin slides are strongly elongated, unlike those in magmatic olivine. The ratio of the length of the crystal to its width is 5:1, with all elongated grains generally parallel to each other, forming a parallel prismatic texture. Similar forms of olivine in bronzite-olivine rocks of the Borzovsk ore deposit were mentioned by V.S. Koptev-Dvornikov and Ye.A. Kuznetsov [13], who came to the conclusion that the olivine was not a magmatic mineral.

The angle of optic axes for olivine ranges from $+82^{\circ}$ to $+88^{\circ}$, occasionally with the negative optic sign for the same value of the angle. The birefringence ranges from 0.032 to 0.036. Judging from these constants, and according to the Poldervaart diagram [23], the olivine composition ranges from forsterite to chrysolite, with 20% the maximum content of fayalite. These data do not quite agree with those of chemical analyses.

If the Al_2O_3, Cr_2O_3, MgO and FeO of dunite be recomputed for chrome-spinelids; and water with corresponding amounts of MgO and SiO_2 -- for serpentine -- arriving in this way at the approximate composition of olivine -- we obtain the following formulas:

Sample 73—1:91.07 $Mg_2SiO_4 \cdot 8.93\ Fe_2SiO_4$

Sample 4—2:94.2 $Mg_2SiO_4 \cdot 5\ Fe_2SiO_4 \cdot 0.8\ CaSiO_4$

The olivine has been but slightly serpentinized. Even chrysolite, which is commonly developed on olivine, is not present in all thin slides. When present, it occurs in fine interlacing veinlets, forming a looped pattern familiar for olivine. It is of interest that dark-colored dunite (dark-green to nearly black) exhibits a distinctly green chrysotile in thin sections, while yellow dunite carries a pale yellow chrysotile with a barely perceptible pleochroism.

Microscopic study of all pyroxene-bearing rocks has established that the pyroxene was first present in large crystals, subsequently shattered to some extent, and replaced with secondary minerals, mostly talc and olivine; and partly, but very seldom, with tremolite. The shattering, bending and microfaulting are the features of the dunite pyroxene. Its replacement with secondary minerals takes place both along the fractures and the grain periphery, thus creating the effect of large pyroxene crystals being broken to bits. The latter usually look like typical relicts enveloped in fine-scaled talc and fine, generally wedge-like olivine grains of a later generation (Fig. 2). Talc commonly makes narrow fringes on the pyroxene relicts, separating them from olivine. Pyroxene has been directly replaced by olivine in many places; in those instances, no perceptible amount of talc is

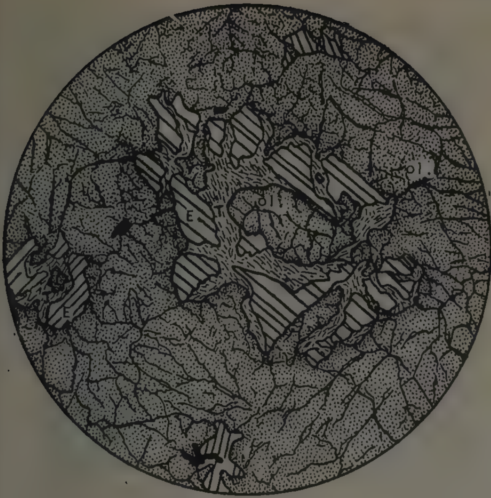

FIGURE 2. Enstatite dunite.

Ol -- olivine; E -- enstatite; T -- talc; black -- chrome-spinelids.

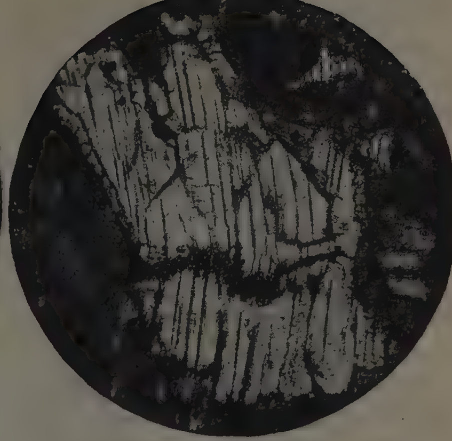

FIGURE 3. Enstatite dunite, with a shattered enstatite grain in the center (white); portions of enstatite grain, on top, are separated by wedge-like olivine grains. Crossed Nicols, 32X.

Table 1

Components	1		2		3	
	% by weight	Molecular amount	% by weight	Molec. amount	% by weight	Molec. amount
SiO_2	40,76	678	38,32	638	53,2	886
TiO_2	None	—	0,40	5	0,10	1
Al_2O_3	2,71	26	2,81	27	2,22	22
Fe_2O_3	0,59	4	2,05	12	3,01	20
Cr_2O_3	0,25	2	None	—	Not det.	—
FeO	7,18	100	5,6	78	4,33	58
MgO	45,32	1124	48,76	1209	33,60	833
CaO	0,20	5	0,40	7	1,79	32
Na_2O+K_2O	None	—	None	—	Traces	—
H_2O	0,56	31	0,30	16	—	—
Loss in heating	1,96		1,62	—	1,48	—
Sum	99,53	—	100,26	—	99,73	—

1 -- dunite with about 10% enstatite content (specimens 73-1); 2 -- dunite (spec. 4-2); 3 -- pyroxenite.

NOTE: Comma represents decimal point.

present on the boundary between pyroxene grains and secondary olivine (Fig. 3).

The pyroxene relicts themselves are fresh and quite colorless, commonly with a fine twinning striation. They usually exhibit direct extinction, sometimes with deviations of 4 to 8°, seldom up to 14°; birefringence, 0.010 to -0.011; the optic angle, from $+78^\circ$ to $+84^\circ$, with only two instances of $+66^\circ$. Thus, in its optical properties, and according to the Poldervaart diagram [26], this pyroxene may be enstatite with an orthoferrosilite content of 7 to 10%. A nearly iron-free enstatite ($2V = 66^\circ$) is much less common.

Chemical analyses give about the same composition for pyroxene. Thus, recomputation of the chemical analyses for spec. 73-1 gave the following formula for pyroxene: $94.4\ MgSiO_3 \cdot 5.6\ FeSiO_3$.

Chrome-spinelids are present in both fairly large and small grains, rounded and irregular; less frequently in well-defined crystals usually included in large olivine grains. In thin sections, the central parts of grains are orange-brown, with black non-transparent fringes along their periphery.

b) Pyroxenite and Related Banded, Ptygmatic and Other Heterogenous Textures of Ultrabasics

As previously noted, pyroxenites are not developed by themselves, but form instead parallel vein-like bands, lenses, folded "veins" and streaky inclusions in dunite.

Banded ultrabasics are widely developed along the southeastern slope of the Range, especially on the headwater plateau of Malyy Abdyr River, where they occur both in outcrops and in loose pieces. Two types of ultrabasics have been differentiated, according to the ratio of rock components of the bands, and according to their mineral composition. The first is characterized macroscopically by the presence of parallel olivine veins in strongly talcose pyroxenite, the most numerous of which are those colored ash gray, 4 to 10 cm thick, or more. They form a sort of framework for a system of vein-like bands of quite fresh olivine, 0.5 to 4 cm thick. On an uneven break, the band boundaries appear straight and sharp. A better picture is obtained, however, on a polished surface. As appears in Figure 4, the contact outlines are very complex. Numerous apohyses with thin branches shoot out of dark olivine bands and into the gray talcose pyroxene. The right band in the picture exhibits numerous gray spots -- talcose pyroxenite within the olivine bands.

As determined under the microscope, the mineral composition of light-colored bands is as follows: talc, 60 to 70%; rhombic pyroxene, 20 to 25%; with impurities of olivine, chrome-spinelids, and occasional tremolite. The pyroxene composition, determined by optical properties, is in no way different from that of the dunite pyroxene. The pyroxene occurs in relicts in the fine-scaled talc aggregate with fine grains of olivine. Its replacement with talc and olivine takes place both on the grain periphery and in the cleavage fractures, with systems of parallel microscopic veinlets originating in the fractures. These veinlets commonly exhibit a zonal structure. There is a trace of the cleavage fracture, in the middle of the veinlet, with a zone of fine-grained olivine on either side, followed by a talc zone of similar or greater thickness on the pyroxene side, and finally by unreplaced pyroxene (Fig. 5).

The olivine is quite fresh, unserpentinized, occurring in small grains usually segregated

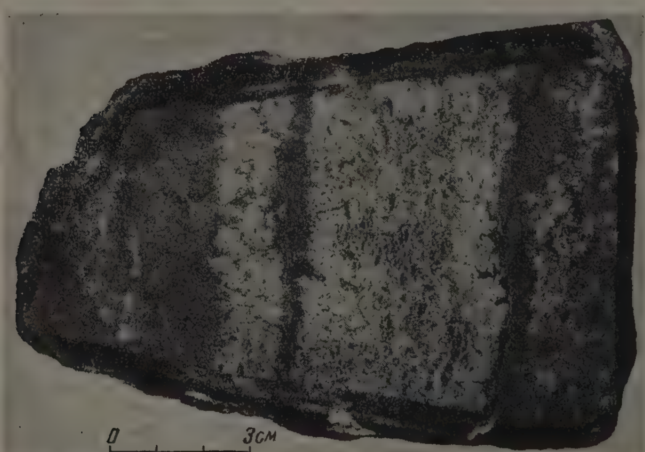

FIGURE 4. Polished section of banded ultrabasic rock.

Black bands are of olivine; light-gray, of talc and relicts of enstatite.

among the pyroxene relicts, with vein-like offshoots into the pyroxene crystals. Its composition, judging from the optic angle (+84°) corresponds to nearly pure forsterite of diagram in paper [20].

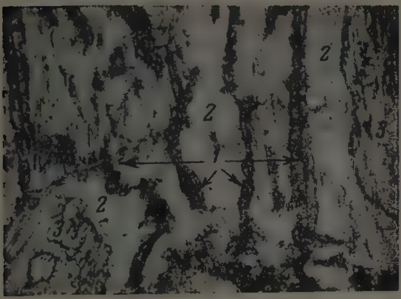

FIGURE 5. Photomicrograph of a light band of banded ultrabasic rock of Figure 4.

1 -- olivine veins; 2 -- talc; 3 -- relicts of pyroxene. Single Nicol; magnification 15X.

The black bands are almost pure olivine, with impurities of chrome-spinelids, pyroxene, and talc. The pyroxene, as determined by its optical properties, is the same as in gray bands, and occurs in relicts of small irregular grains. Adjacent pyroxene grains, separated by talc and olivine, frequently have the same orientation, suggesting that they belong to a single large crystal. Thus, the pyroxene-olivine ratio in black bands is the same as in the gray, with only the quantitative mineral content different. Talc forms fringes on the pyroxene relicts, separating pyroxene from olivine. However, direct replacement of pyroxene by olivine is not uncommon, in which case there is no perceptible amount of talc formed on the boundary between pyroxene and olivine grains.

Tremolite is rare. A single thin section displayed a tremolite crystal at the contact between the gray and black bands. The crystal has been cut by the plane of the section, normal to its prism faces, and is enveloped on three sides by an olivine grain, which makes it idiomorphic with the olivine. Clean-cut idiomorphism of tremolite with olivine has been noted by A.N. Zavaritskiy and O. Baklund, in Uralian rocks of a similar composition. For this reason, O. Baklund regarded tremolite as a primary mineral formed before olivine.

The second type of banding is represented by a system of pyroxene "veinlets," usually dark-brown or black, in massively crystalline greenish gray dunite. These "veinlets" are 2 to 6 cm wide, sometimes more, seldom attaining 30 to 40 cm. Scattered in a parallel orientation near each other, they create the banding effect in dunite. Pyroxene being more resistive to weathering than olivine, its "veinlets" stand out 1 to 2 cm high on weathered surfaces, giving them a corrugated aspect. The most interesting feature of this banding is that the pyroxene "veinlets" are often broken, as if constricted by olivine aggregate. The surface of such chunks shows only fragments of pyroxene "veinlets" which occur in the dunite body as tabular "xenoliths," while the rock as a whole looks like a typical replacement breccia.

Figure 6 is a photograph of an essentially olivine block, with pyroxenite occurring in tabular "xenoliths" and "veins," more or less replaced with olivine aggregate. Some of these "veins" extend across the entire block, thus preserving their banded aspect; the others have been wholly replaced, displaying nothing but isolated pyroxene crystals, in a chain through the dunite body. In some dunite blocks, the replaced pyroxenite "veinlets" look like chains of rounded pyroxenite fragments. On weathered surfaces, these "fragments" stand out in sharp relief above the olivine rock, pebble-like (Fig. 7).

These and similar forms could not have originated other than from replacement of pyroxenite "veins," not only at the contact with the enclosing rock but in the transverse fractures cutting these "veins."

Ptygmatic textures are morphologically and genetically similar to those of granitized rocks and migmatites, for which N.G. Sudovikov [18] proposes the name, pseudoptygmatic. In our rocks, they present folded veins in blocks of massive non-schistose dunite. These veins are of the same coarse-grained pyroxenite as the tabular "xenoliths" in dunite. The comparatively large folds, 50 to 70 cm high, are locally complicated by secondary folds, up to several centimeters high (Figs. 8, 9). These folded "veins" are usually seen on one or two opposite sides of a block, appearing on the other side as a system of parallel "veinlets." In other words, the trend of folds, in a cross-section, produces the above-described banding. In the apices of folds, the pyroxenite "veins" are usually thickened to 10 to 15 cm, which is several times more than their thickness in the wings.

Macroscopically, the pyroxenite forms dark gray to almost black or greenish brown rocks, usually medium- to coarsely-crystalline. They are always fresh, monomineralic, consisting of isometric to slightly elongated

FIGURE 6. Processed surface of banded ultrabasic rock.

1 -- xenolith-like tablets of pyroxenite in dunite; 2 -- pyroxenite bands, nearly fully replaced by olivine aggregate; 3 -- chain-like distribution of pyroxene grains in dunite.

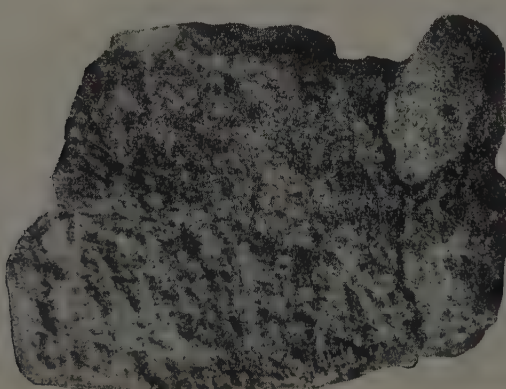

FIGURE 7. Dunite specimen with a pebble-like pyroxenite inclusion.

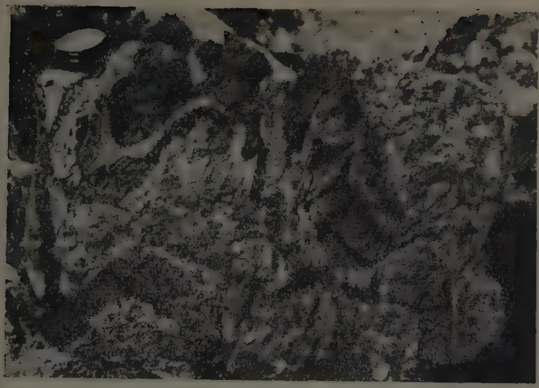

FIGURE 8. Folded "veins" of pyroxenite (black) in dunite (pegmatitic structure).

pyroxene tablets, 5 to 10 mm thick, lustrous on cleavage planes. No instances have been observed of a decrease in grain size in "veins," at the contact with the enclosing olivine rock. There are pyroxenite "veinlets," 3 to 4 cm thick, built up of 2 to 3 cm crystals, which take up nearly its entire thickness. Similarly large pyroxene crystals in thin veins have been noted from the Polar Ural dunites. A.N. Zavaritskiy stated on the subject of their origin, "Such crystals could not have been formed other than by replacement" [5]. We accept that as an explanation of the origin of the pyroxenite rocks in question, including the folded ones.

Microscopic study has revealed that pyroxenites which form vein-like bands, tabular "xenoliths," and folded "veins" in dunite, carry besides the pyroxene impurities of talc, olivine, and isolated grains of chrome-spinelids. Judging from the optic angle (+84°), and

FIGURE 9. Processed surface of a dunite-pyroxenite block with banded-ptygmatitic texture.

The compass rests in the middle of a pyroxenite fold. Below there are two thin folded "veinlets" of pyroxenite, nearly fully replaced with olivine aggregate and represented, accordingly, by a chain of rounded pyroxenite fragments (1); 2 -- thin bands of pyroxenite in dunite; 3 -- wide, vein-like band of pyroxenite.

other optical properties, this pyroxene is enstatite and is in no way different from that of banded varieties of the first type. The olivine content in pyroxenite ranges from 0 to 10^{o}. Thus, in terms of mineral composition, they may be defined as enstatite and olivine enstatite.

Structurally, the enstatite occurs in coarse, strongly cataclastic crystals, nearly everywhere exhibiting wavy extinction, bending of the cleavage fractures, and noticeable breaks and microfaults. Small olivine grains, piled together, are common among the enstatite crystals. Olivine, alone or in veins with talc, commonly penetrates along the cleavage planes, deep into enstatite grains, or else is developed in wedge-like aggregates in breaks splitting the enstatite crystals.

The microscopic study shows, then, that all olivine and pyroxene-bearing rocks, enstatite dunite, harzburgite, or olivine pyroxenite, are marked by structural features suggesting the replacement of pyroxene by olivine, with talc frequently developed on pyroxene, as a secondary mineral. Because of this relationship, pyroxene occurs in relicts within enstatite dunite and harzburgite, while olivine of pyroxenite forms veinlets and wedge-like aggregates penetrating the pyroxene crystals. Pyroxene grains are obviously cataclastic, while olivine is almost undisturbed.

Thus, the microscopic study, together with the mineral composition and the relationship between minerals, presents convincing evidence that, in all instances of parallel occurrence of pyroxenite and dunite, pyroxenite is the earlier formation, and that all textural varieties differ from each other only morphologically; genetically, they are products of the same processes.

A perusal of the literature on ultrabasics of other regions reveals that our pyroxenites and associated banded and streaky textures are not unique. The only thing we did not find was a mention of ptygmatic textures.

It is pertinent to mention here, however briefly, some of the publications on pyroxenites in association with serpentinite and dunite. First of all, there are N.I. Bezborod'ko's mention (quoted from [12, 17]) of those vein- and druse-like sectors of a bronzite rock in the Maikop serpentine, described by him as "kernels" and supposed to be remnants of ancestral serpentinite. The error of Bezborod'ko's view that serpentinites of that region originated from a pyroxene intrusion was subsequently proven by a number of investigators. V.N. Lodochnikov [12] came to the conclusion that "segments of the monomineral bronzite rock observed, according to N.I. Bezborod'ko, in the Maikop serpentine, cannot be accepted as remnants of an extrusive rock." He also noted that pyroxenites were more resistant to serpentinization than a nearly pure olivine rock -- a fact confirmed by many investigators in many instances. If this is true, there is every reason to believe that when our dunites with pyroxenite in tabular "xenoliths," streaky inclusions, and other bodies, undergo serpentinization -- as they eventually must -- most of these pyroxenites will be preserved in the serpentine body, as is the case of the Maikop serpentines. Thus, these pyroxenites will become relicts for the second time.

Small bodies or pyroxenite, similar to ours, are fairly common in serpentine and dunite. They were described by B.P. Krotov [10], P.M. Tatarinov [20], A.N. Zavaritskiy [6] and other students of the Urals. All observers of small pyroxenite bodies in ultrabasics tentatively identified them as "schlieren" (indistinct streaky inclusions). According to N.D. Sobolev [17], pyroxenites of the Greater Caucasus are represented by such streaky and vein inclusions. A.N. Aleshkov [2] observed pyroxene xenoliths in dunite, in a number of the polar Ural localities, and noted the presence of dunite veins in pyroxenite. Many authors, such as V.N. Lodochnikov, A.N. Zavaritskiy, V.S. Koptev-Dvornikov, Ye. A. Kuznetsov, V.I. Luchitskiy, A.A. Kadenskiy and others, discussed the origin of pyroxenites which they observed among serpentinite and other ultrabasic rocks. V.S. Koptev-Dvornikov, Ye. A. Kuznetsov, V.N. Lodochnikov and others believe them to be metasomatic, solely because they are coarsely crystalline, with grain size up to one centimeter and often much larger-- this despite the small dimensions of their bodies. Lodochnikov has compiled and systematized voluminous data to support his contention that "rhombic pyroxene in large accumulations and crystals is as typical a hydrothermal mineral in ultrabasics as a coarse pegmatitic microcline is in granite." ([12], page 234.)

According to Lodochnikov, rhombic pyroxene is formed from primary olivine of peridotite, as follows: olivine + $SiO_2 \rightarrow$ rhombic pyroxene. Kadenskiy [7], in his study of the origin of the Ekhresku enstatite range, the Caucasus, arrived at the same conclusion.

On the basis of the above data, it may be stated that the formation of pyroxenite under conditions of a great influx of silica is possible not only at the expense of olivine-rich rocks, but of any rock with the MgO/SiO_2 ratio greater than in the enstatite molecule.

THE THEORY OF MAGNESIUM METASOMATISM IN SERPENTINES

As pointed out above, the structural and textural features of dunites and pyroxenites of the Borus Range, and their relationship with each other and with the enclosing serpentinite, are utterly inexplicable in the light of the magmatic melt crystallization concept. There is every reason to associate the formation of dunite and pyroxenite with metasomatic processes, as indicated by much of the above data. However, the true effect of supersaturated magnesium solutions is reflected, in addition, in the structural diversity of metasomatic olivine veins in serpentinite housing the dunite bodies. On the whole, the metasomatic nature of some of these veins is unquestionable. Figures 10 and 11 are photomicrographs of two olivine veinlets. It is readily seen that both of them, together, graphically illustrate the history of formation of one of them (Fig. 11), as an end product. The first vein is not strictly olivine. The fissure has been filled with a nearly isotropic serpentine-like substance of an unknown composition. An olivine fringe is present on both sides, apparently replacing the adjacent serpentine rock. The second veinlet consists of a similar olivine fringe; in addition, the fissure itself has been filled with olivine. There is every reason to regard this and similar veinlets as a manifestation of the most elementary near-fracture replacement.

It is well known that the main factors controlling the mineral-forming process in metasomatism are the thermodynamic environment and the mobility of components. D. S. Korzhinskiy [8] believes that, in addition, the concentration gradient is as important a metasomatic factor, if not more so, than the others. It readily follows from the chemical composition of olivine and serpentine that the solutions were supersaturated with magnesium. In our case, the magnesium concentration in the pore-filling solution decreased from the fissure to the enclosing rock, while the silica concentration decreased toward the fissure. The resulting difference in the concentration of these main components of the system led to diffusion processes. Inasmuch as the influx of magnesium resulted in its excess in the serpentine pore-filling solution, in relation to the magnesium content of serpentine, the serpentine molecule became unstable, under proper physical conditions, and dissociated as follows:

$$2H_4Mg_3Si_2O_9 \rightarrow 3Mg_2SiO_4 + SiO_2 + 4H_2O.$$

Some of the liberated silica immediately united with the excess magnesium in solution,

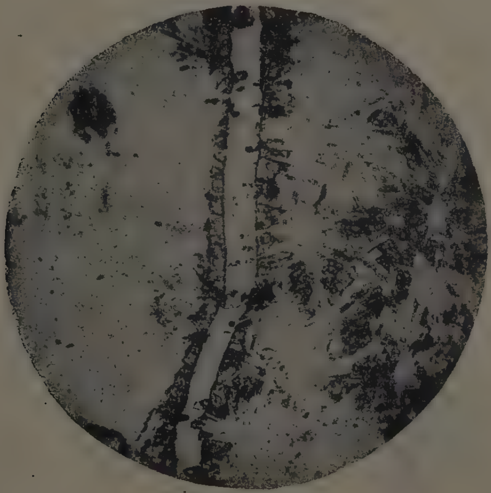

FIGURE 10. Serpentinized ultrabasic with vein of a serpentine-like mineral.

Vein is fringed with olivine crystals. Single Nicol; magnification, 40X.

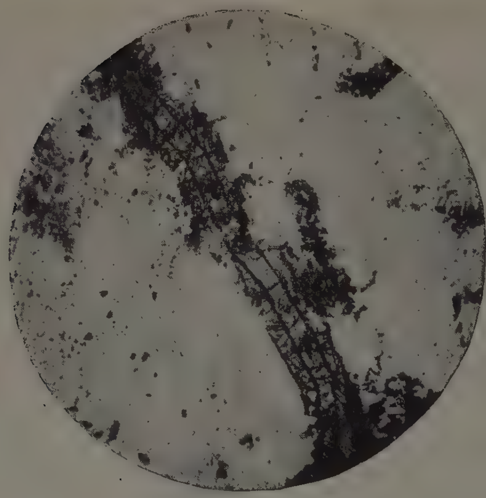

FIGURE 11. Olivine veinlet in the serpentinized ultrabasic of Figure 10.

Single Nicol; 40X.

to form the olivine molecule $2MgO + SiO_2 \rightarrow Mg_2SiO_4$.

The olivine so formed occupied the space vacated by silica and water in the serpentine molecule. The balance of silica, because of the decrease in its concentration in the direction of the circulating solution, migrated to the fissure, together with water. Consequently, the infiltrating solution was enriched in silica, until its concentration attained the minimum value necessary for the formation of olivine.

Thus, a parallel system of olivine veinlets should be formed first in strongly fractured serpentinite. As the process goes on, the veinlets should come together to form a massive vein-like dunite body.

It has been pointed out that vein-like dunite bodies are common at the serpentinite contact with Dunite Body One. Some of them are somewhat slaty. Their thickness ranges from 1 to 4 mm, with joint planes 2 to 5 mm apart. The total lack of pyroxene in them, and the incipient slate cleavage as expressed in thin partings coinciding with the strike, in the absence of any traces of schistosity and cataclastic effects, make it possible to relate the origin of dunite and the parallel partings to the same processes of near-fracture diffusion metasomatism in serpentine which were responsible for the microscopic veinlets of olivine. The olivine veinlet in Figure 11 exhibits plainly the boundaries of the original fracture, now separating the olivine replacing serpentine from that crystallized in the fracture, out of the infiltrating solution. The sharp outline of these boundaries undoubtedly suggests the lack of fast ties between the two generations of olivine. This, in the subsequent cooling, should lead to the formation of a system of parallel fissures. Thus, a single fracture in the original serpentinite becomes two, in the dunite, as a result of the near-fracture metasomatism.

Experimental study has shown [4] that olivine is readily synthesized by the hydrothermal method, at a temperature of 500°C and above, and is stable in the presence of water vapor, up to 430°C. According to the same data, serpentine can be formed at a temperature of about 500°C. Thus, the temperature conditions coincide for both minerals at 500°C. Of course, the experimental results obtained from a fairly simple system may not be fully applicable to natural processes which are more complex and take place in hydrothermal solutions carrying various amounts of other components, especially volatile ones. The latter, as is well-known, lower rather than raise the crystallization temperature of minerals. It follows that if olivine and serpentine have a narrow range in temperature of formation, under laboratory conditions, this range may be considerably wider under natural conditions. In that case, the main

factor determining the sequence of precipitation for minerals will not be the thermodynamic environment (which is a necessary condition only) but rather the concentration of main components, silica and magnesium, in the solution.

A rise in the silica concentration in the infiltrating solution should lead first to precipitation of serpentine. Only when the excess silica is expended, will olivine crystallize at the same temperature and pressure. Antigorite, which is formed in such processes, will be found in poikilitic growth in olivine. This relationship between olivine and antigorite was noted by E. Weinschenck and B.P. Krotov, in their time. Krotov convincingly proved that antigorite is a primary mineral, illustrating his thesis with photographs [9].

As regards those serpentine segments characterized by little fracturing and a more or less uniform development of porosity, the replacement there takes place by the diffusion of elements in the pore-filling solution. Because of dissociation of the serpentine molecule, the silica concentration in the pore-filling solution will rise; the solution will be less undersaturated with silica and more diluted, because water will have passed from the serpentine molecule to the intergranular space. Further dissociation of the serpentine molecule will be slowed down. All this leads to the following conclusions:

1. In the absence of open fissures in serpentine in the process of replacement, the desilification front cannot overtake too much, or get away from, the basification front. Therefore, a gradual increase in the metasomatically originated olivine in serpentinite is not to be expected. The contact must be fairly sharp. Gradual changes are possible not in a monomineral rock such as serpentinite but apparently in a rock with a more complex mineral composition.

2. A higher silica concentration in a pore-filling solution brings about a concentration change in the active solution. As a consequence, a mineral with a higher silica concentration will crystallize first. This means enstatite rather than olivine, because the silica content in enstatite is twice as high as in olivine.

Since the replacement takes place mostly with a constant volume, the silica in solution is very mobile with relation to magnesium. Where fractures are present, siliceous solutions will use them as means of escape to higher horizons. Thus, leaching of some

FIGURE 12. Diagram of the development of magnesium metasomatism in serpentinite.

1A -- fractures; 1B -- pyroxene veins in serpentinite; 2 -- olivine-pyroxene rock; 3 -- dunite; 4 -- pyroxenite; 5 -- tabular "xenoliths" of pyroxenite in dunite and olivine-pyroxene rock. Arrows point in the direction of the silica movement from the main basification front; size of arrows designates its velocity.

silica will be accomplished by the infiltrating solution. The appearance of enstatite veins should be regarded as a result of a chemical reaction between hard rocks comparatively rich in magnesium (serpentinite and possibly some others), and siliceous solutions from the desilicification of the same serpentinite, at greater depths (Fig. 12). The process of enstatite formation can be presented as follows:

$$H_4Mg_3Si_2O_9 + SiO_2 \rightarrow 3MgSiO_3 + 2H_2O.$$

Thus, enstatite veins mark the "footprints" of silica going away from the main basification front. Accordingly, metasomatism cannot lead to the formation of a single replacement rock, wherever the substance is brought by carrier solutions, without bringing about the appearance of another replacement rock at the places of migration of the expelled substance.

The form of veins was determined by that of the fissures -- the circulation channels of solutions. Folded rocks produced folded veins. The association of folded pyroxenite veins in contact with dunite with enclosing rocks suggests that these veins originated in partial replacement of folded emplacing rocks leaving enstatite. This replacement was selective along definite directions which were not only readily accessible to the solutions but also marked by a composition most favorable for replacement.

The ascending metasomatizing solutions encountered and flushed olivine-pyroxene rocks approximately corresponding to harzburgite in composition. As the main basification front ascended, these solutions encountered serpentine with enstatite veins (Fig. 12). With enstatite being stable in a wide temperature range, the replacement of these rocks was highly selective. Serpentinite was replaced while enstatite veins persisted as relicts and became part of a newly-originated olivine-pyroxene or olivine rock. A parallel system of such relict veins in dunite is what determines the appearance of the above-mentioned banded and ptygmatic textures.

The replacement processes, however, do not terminate there, provided the supply of metasomatizing solutions keeps up. As long as the MgO/SiO_2 ratio in enstatite is only a half of that in olivine or, in other words, enstatite carries twice as much silica, and as long as the solution maintains its magnesium composition, enstatite will be the less stable under given thermodynamic conditions, even with the same general stability for it and olivine. The dissociation of enstatite, like that of serpentine, first leads to a rise in the silica concentration in the pore-filling solution. As a result, talc is produced instead of olivine. Only in the subsequent reaction with the solution is talc replaced by second generation olivine.

Thus, enstatite fully disappears in the lower horizons where it is subjected to a protracted action of the magnesium-bearing solution. Olivine-pyroxene rocks turn to dunite, while the relict enstatite veins are broken into tabular "xenoliths." The emergence of pyroxenite may be anticipated in upper horizons, at the terminal metasomatic stage, with an olivine-pyroxene rock (harzburgite) persisting below; and dunite, still lower. This concept is in full agreement with D. S. Korzhinskiy's theory of metasomatic zonation according to which there should be a column of metasomatic zones, growing by virtue of rear zones catching up with the forward ones, wherein one metasomatic zone is replaced by another, of a simpler mineral composition, until a monomineral rock has been formed.

SUMMARY

1. Dunite and genetically and spatially associated pyroxenite are younger than serpentinized ultrabasics of the Borus massif.

2. Dunite and pyroxenite are qualitatively different products of two interconnected stages of the same metasomatic process, each representing a change in the quantitative ratios of reactive components MgO and SiO_2 at different depths of the rock being replaced.

3. Products of the first metasomatic stage -- simple and folded pyroxenite veins -- have been formed in a chemical reaction between magnesium-rich rocks with siliceous solutions ascending along fractures. These siliceous solutions originated as an unavoidable result of the leaching of excess silica at depth, where essentially olivine rocks -- dunite -- were formed as a result of an influx of magnesium.

4. The replacement was highly selective in both metasomatic stages. Pyroxene veins, which originated at a given horizon, prior to the advent of dunite, were only partly replaced by the dunite, as the main replacement front advanced. Most of them persisted as relicts -- tabular "xenoliths," streaky inclusions and other forms.

5. The occurrence of ptygmatic textures at the albite-epidote-chlorite gneiss contacts suggests a replacement of already folded rocks. Thus, dunite and pyroxenite must have been formed either after the main folding phase or else during its terminal stages.

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Received January 21, 1958

THE STRUCTURE AND AGE OF THE SOVGAVAN' FORMATION OF THE SIKHOTE-ALIN, NORTH OF THE KHUTSIN HARBOR MERIDIAN[1]

by

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The Sovgavan' formation includes the youngest extrusives of the Sikhote-Alin. They form a gently dissected plateau, several hundred to several thousand square kilometers in area, in the basins of Zeva, Bikin, Takhobe, Nel'ma, Issima, Koppa Rivers, and in the Sovgavan' area (Fig. 1). This is a fairly homogeneous sequence of chiefly finely porous basalt, locally coarsely porous and vesicular, and occasionally dense, mostly gray to dark gray, locally black. The dense dark gray to black varieties are developed chiefly at the base. Their most complete sections were studied by A.B. Razzhivin, in the Soviet Harbor area where they have been penetrated by drilling, down to 270 m.

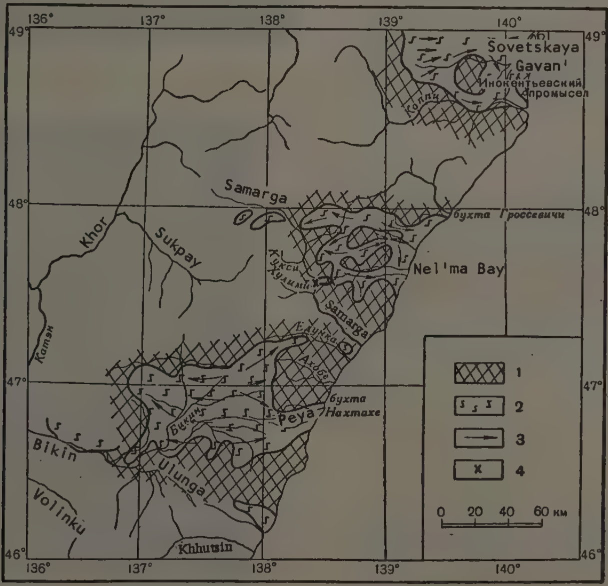

FIGURE 1. Paleogeographic map of the Sovgavan' basalt lava flows.

1 -- high points of ancient terrain, bypassed by basalt lava; 2 -- basalt plateaus; 3 -- direction of lava flow; 4 -- gap.

Some of the exposures, whose structure

[1]Stroyeniye i vozrast Sovgavan'skoy svity Sikhote-Alina k severu ot shiroty Bukhty Kkhutsin.

and composition are typical of this formation, are described below.

The following section was described from the Tatar Straits shore cliffs, 5 km southwest of Krestovozdvizhenskiy Point (from the bottom upward):

1. Basalt, black, dense, very tough, with a well-defined columnar jointing of 4 to 5 faced prisms (Fig. 2, 3).

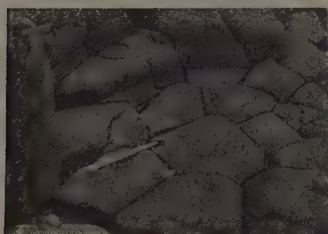

FIGURE 2. 4 to 5 sided columnar jointing in basalt, seen from above.

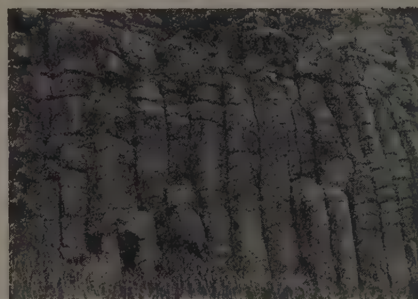

FIGURE 3. Same as Fig. 2; side view.

2. Basalt differing from the preceding type in its fine porosity and dark gray color. The contact between the two is fairly uneven, but is not a boundary between two lava sheets. Rather, the upper part of a lava flow was enriched here with escaping gases. This has led to the formation of vesicles and vugs and, consequently to the change in color and strength. Thickness, 3 m.

3. Gray, finely porous basalt -- and locally black, dense basalt -- overlain by dark gray basalt, not as porous and with block parting, and occasionally bubbly. Thickness, 2 m.

4. The section is crowned by bubbly gray basalt. Exposed thickness, 10 to 15 m.

A well-defined stratification in the basalt sequence is present in the left bank cliff along the Nel'ma-Lugovaya harbor. It is due to the presence of brown-red rocks and weathered crusts.

1) At the very base of the cliff, there are gray, finely porous, very tough basalts with lenses and intercalations of bubbly varieties (Fig. 4). Locally, they have a honey-combed weathered surface (Fig. 5) and change upward to vuggy or strongly bubbly varieties, 0.5 to 1.0 m thick. The overall thickness, in this locality, about 50 m.

The lava blanket is capped by a weathered crust of red-brown, loose eluvium, 1 m thick.

2) This uneven weathered crust is overlain by gray, finely porous basalt, changing upward to red-brown bubbly lava, as much as 2 m thick. Overall thickness, 10 m.

FIGURE 4. Finely porous gray basalt with thin intercalations of bubbly varieties.

3. A basal sheet, similar to the second unit. Thickness, 8 m.

Higher up the shore incline, both finely-porous gray and bubbly red-brown basalts are seen in blocks. Their total thickness here is

FIGURE 5. Honeycombed weathered surface of gray, finely porous basalt.

about 150 m. A stratified succession of basalt sheets -- very rarely of andesite basalt -- has been observed in exposures and boreholes, in the Sovgavan' area. These basalt sheets, 2 to 10 m thick, are separated by bedding surfaces or weathered crusts.

These sections graphically illustrate the periodic volcanism responsible for the formation of the basalt sequence. Judging from the variable thickness of the weathered crust, the periods between different lava flows were of different durations.

There are more of such sequences, but there is no need of citing them, because they are all alike in composition, structure and textural features.

A distinctive feature of the Tertiary volcanism is the total lack of pyroclastics. Lava breccias are exposed in a small ancient gap near the mouths of the Amakhe and Takhobe Rivers. They contain fragments of the underlying Samarginsk Paleocene dacite, which accounts for a small percentage of the total, and of basalt. The basalt fragments apparently are products of disintegration of the first basalt flow, subsequently cemented by later lava flows.

Strewn over the basalt plateaus are rounded blocks, up to 1 m in diameter, seldom larger (the Sovgavan' area, etc.). We concur with A.N. Zavaritskiy [4] who regards the blocky lava as having been formed as a result of breaking up of the upper layer during the process of solidification. Only isolated blocks, rounded off by weathering, have persisted into the present.

The basalt sheets lie either horizontally or inclined, according to their initial dip and the relief of the surface on which they were deposited. Faults have not been seen in the basalt plateaus; it is not impossible, however, that they are present in the lava flow zones.

The total lack of pyroclastic formations in the basalt sequence, as well as the lack of any evidence of volcanic centers, suggests the predominance of a fracture-type lava extrusion.

Field data throughout the basalt plateaus suggest that lavas were poured into low places of the terrain, over rocks of different ages, including the Neogene Kizinsk extrusives. This is supported by the following data:

1. Sharp differences in elevation of the basalt base (from 0 to 1,000 m), over small areas, with the base rising away from the present base of erosion (Sovgavan', Nel'ma, etc.).

2. The presence of basalt in river terraces, 15 to 20 m high.

3. The surface slope of the basalt plateaus toward the present base of erosion.

4. The basalt lavas flowed around the contemporaneous high points, such as the Konskaya Golova (Horse Head) in the upper course of Zeva River; Khabinskiye Belki, and Chipali ranges (Fig. 1).

Thus, it appears from an analysis of the above data and from the configuration of the basalt plateaus that the lavas descended down the valleys of contemporaneous rivers, toward the Sea of Japan and the Tatar Straits. In addition, it filled tectonic depressions, such as the Upper Bikinskaya, made up of Oligocene-Miocene coal measures.

The thickness of basalt depended on the relief. As long as it filled the relief lows, it appears that its thickness should increase from the periphery toward the center of these plateaus, and with the direction of the flow.

The Sovgavan' formation is marked by a fairly uniform ratio between its component parts, which usually are basalts, with occasional andesite-basalt or plagiobasalt.

The basalt usually is equigranular, very seldom porphyritic, with porphyritic inclusions usually carrying olivine in idiomorphic crystals, inconspicuous in the groundmass. Plagioclase also is very rare in the inclusions, noticeable only under the microscope, where it, too, is inconspicuous. These basalts are most often evenly crystallized, with a doleritic, ophitic or poikiloophitic texture (Fig. 6). In the upper parts of the flows, the rock structure is intersertal, often hyalointersertal, marked with a sizable content of nearly black, highly ferruginous volcanic glass with growths

FIGURE 6. Coarse pyroxene crystal pierced by fine plagioclase growth, in the Sovgavan' basalt.

28 X; Nicols crossed.

FIGURE 7. Micropegmatite in fully crystallized basalt.

42 X; Nicols crossed.

(leists) of plagioclase and rare grains of pyroxene and olivine.

The basalts consist of labradorite No. 54-60 and pyroxene-augite represented by a ferruginous variety, C: γ = 40 — 43°; γ = 1.720; α = 1.695; $\gamma - \alpha$ = 0.25; 2V, about 50°. There are occasional idiomorphic crystals of olivine, commonly replaced with iddingsite and fringed with brown iron oxide. No hypersthene has been found in the Sovgavan' basalt.

Near the centers of basalt outpourings, where the crystallization conditions were the most favorable, the interstices between plagioclase crystals are commonly filled with micropegmatite of quartz and acid plagioclase growths (Fig. 7). Microscopically, such rocks are very similar to microdiorite. These basalts have been observed along the Mitsuyevsk spring, in the interior of the Uysk highland caldron (the Sovgavan').

The Sovgavan' basalt usually is finely porous, bubbly and vuggy at the top, which points to a lava saturated with gas and vapor and therefore mobile. The pores are locally filled with zeolites and calcite.

The silicate analyses of the Sovgavan' basalt reveals a fairly homogeneous composition, approaching the average quartz basalt (after R. Daly), and only to a small extent the andesite-basalts and the average plateau basalt (after R. Daly).

There has been a difference of opinion as to the age of the Sikhote-Alin basalt. Ya. S. Edel'shteyn [9], on the basis of their resting directly upon ancient alluvium of Bikin River, believed them to be Quaternary. N.A. Belyayevskiy [1], M.A. Favorskaya [7], Yu. F. Chemekov [8], G.S. Ganeshin [3] and others, assigned them to the upper Quaternary.

P.N. Kropotkin [5] and I.I. Bersenev [2] assign a Pliocene age to the plateau basalts, because they rest on the Pliocene Suyfun formation of the southern Maritime Region (Primor'ye), where they carry at their base sedimentary beds with upper Miocene and Pliocene pollen assemblages.

In 1955, this author assigned the Middle Sikhote-Alin basalt to the Neo-Pleistocene, on the following indirect considerations:

1. Plateau basalts are located in river terraces, 15 to 20 m high (Bikin, Takhobe, Zeva, and other river valleys).

2. There are no basalt pebbles in alluvial deposits of Terrace III (15 to 20 m high) on Samarga River, 5 km downstream from the basalt plateau.

3. There are no terraces older than the recent, in valleys which cut the plateau.

4. The basalt plateaus are but slightly dissected, with no tributaries to the valleys

which cut them.

5. The progressive erosion in these valleys has reached only the middle of the plateau, above which the rivers flow almost at the surface (the Zeva, Nakhtakhe, and other rivers).

6. The presence of gaps in the valleys of the Samarga and others, where the slopes have been built up of basalt, almost to the rim; this means that the basalt filled up the valleys almost as deep as they are now.

7. Basalt fills up the valleys in Neogene extrusives (Zeva, Takhobe, and other rivers).

The 1956 data of spore-pollen analysis of alluvial deposits underlying the basalt have a decisive bearing, in our opinion, on the age of the Sikhote-Alin basalt, which is the subject of this paper.

We determined the position of basalt on alluvial deposits in four localities: on the right bank of the Tumnin, 20 km from the mouth; on the right bank of the Bikin, 10 km both above and below Ulunga settlement; in Takhobe harbor, north of the river mouth of the same name; and on the right bank of the Iman, at Kartun settlement. In addition, geologists of the All-Union Geological Institute (VSEGEI) have observed nearly the same situation on the left side of the Bikin valley, 16 km above the mouth of a left tributary of the Zeva, and on the right side of the Zeva valley, 14 km from the mouth.

In a railroad cut on the right side of the Tumnin valley, A.B. Razzhivin and E.K. Dul'kis studied, in 1956, a section of alluvial and volcanic formations which fill an ancient erosional hollow where they rest upon Tertiary extrusives.

V.P. Grichuk identified a spore-pollen assemblage in samples of alluvial deposits collected for that purpose by A.B. Razzhivin.

V.P. Grichuk notes that "only a small amount of pollen and spores (25 to 75) has been found in all samples. Because of the composition of their spectra, with only living genera and species present, these deposits should be assigned to the Pleistocene, although the Upper Pleistocene is not to be ruled out."

Alluvial deposits, 4 to 6 m thick, consisting of sand gravel and pebbles, are located on the right bank of the Bikin, 10 km below Ulunga, where they rest upon disturbed Mesozoic deposits forming shelves 5 to 8 m high.

These alluvial deposits are overlain by basalt sheets, as much as 10 m thick.

In his spore-pollen analysis of our 1956 samples from these alluvial deposits, Grichuk noted that "at the present stage of knowledge of the Cenozoic of the Far East, it is impossible to determine the exact Pleistocene age of these beds."

Pebble beds and sandstone rest upon an uneven Samarginsk dacite surface, forming a shelf, 10 to 12 m high, in the Takhobe harbor area, north of the river mouth of the same name. These beds are covered by basalt, 3 to 4 m thick.

V.P. Grichuk has found about 500 grains of Quaternary pollen and spores in sandstone below the basalt.

Grichuk states, "On the basis of composition of the flora and in the character of spore-pollen spectra, these deposits may be assigned to the Pleistocene." He notes in addition, that "no obvious contamination by modern pollen has been observed in the analysed samples. There is every reason to believe that spores and pollen are in situ."

Similar relationships between basalts on river terraces, as much as 20 m high, were observed by K.M. Khudoley, in the Bikin valley (above the confluence with the Zeva) and in the Zeva valley. A spore-pollen analysis by I.A. Sivertseva, in consultation with M.A. Sedova, also indicates a Quaternary age for alluvial deposits underlying the plateau basalts.

The results of all spore-pollen analyses of alluvial deposits underlying the basalt, and of the analysis of geomorphic features of valleys which cut the plateaus, give sufficient reason for assigning an upper Quaternary age to the basalt. It is not impossible that some of the individual sheets belong to older divisions of the Quaternary.

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Received July 14, 1957

PETROGRAPHIC STUDY OF CLAYS FROM MAIKOP FORMATION OF THE AZERBAYDZHAN CIS-CASPIAN OIL PROVINCE[1]

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Detailed lithologic study of probable oil source beds, and the determination of their diagnostic features, have acquired particular significance and importance in the solution of a number of theoretical and practical problems of petroleum geology [3, 4]. This paper represents that portion of a comprehensive study of the Azerbaydzhan Maikop clays which deals with their occurrence in the Cis-Caspian oil province, i.e., in the foothill zone of northeastern Azerbaydzhan, between the Kiyazinsk sand bar in the southeast and Samur River in the northwest [2, 5, 6].

Facies changes have been observed in the Cis-Caspian Maikop deposits. The northwestern part of this formation is represented by an arenaceous-argillaceous facies, becoming argillaceous in its southeastern part where its thickness has been cut down to 450 m [2] [1, 5, 10, 11].

The most complete section of the arenaceous-argillaceous facies of the Maikop formation has been uncovered by deep and surface drilling in the area of Amirkhanly and Saadan. The Maikop beds rest here monoclinally below overthrust Cretaceous rocks [7]. Lithologically, this section of the Maikop formation is divisible into two subformations: the upper and the lower. Argillaceous rocks account for 60 to 70% of the total, locally as much as 80%.

The mineral composition, origin, and geochemical conditions of the Maikop sedimentary basin have been studied by petrographic and chemical methods; the clays, by the granulometric, optical immersion, thermal, chemical, electron microscopic, organic dye, and other methods [8].

The laboratory samples were obtained from deep and exploratory wells drilled in the area of Amirkhanly, Saadan (central part of the Cis-Caspian province) and Yashma (its southeastern district).

The Results of Study

From granulometric analyses, the clays are very sandy, with the sand content between 35 and 40% in the Saadan section, and 40 to 45% in the Amirkhanly section. At the same time, upper Maikop clays are slightly calcareous, with the maximum lime content 3%, while the Lower Maikop clays are strongly calcareous, with the $CaCO_3$ content ranging from 3 to 55.6%.

The immersion method study of the optical constants of clay particles in oriented aggregates has shown that the value of γ for upper Maikop clays ranges from 1.549 to 1.555, occasionally reaching 1.568; α ranges from 1.510 to 1.550, at times up to 1.559. In lower Maikop clays, γ ranges from 1.562 to 1.572; α, from 1.556 to 1.558.

The microscopic study of thin sections has revealed that the clays have an essentially pelitic, silty-pelitic, and psammo-pelitic structure.

It should be noted that most Maikop clays contain a fair amount of organic matter of a mixed type. Its content, visually determined under the microscope, ranges from 5 to 10%.

Present among terrigenous and syngenetic minerals are quartz, 10 to 15%; feldspars, 3 to 5%; pyrite, up to 20%; muscovite and chlorite, 2 to 5%; carbonates, from 5 to 10%; siderite, up to 3%; organic matter, including bitumens, 5 to 6%, occasionally 10%, and others. Noted among clay minerals are montmorillonite, beidellite, hydromica and kaolinite. The content of rock-forming minerals in thin sections ranges from 30 to 70%.

The light portion of lower Maikop non-

[1] Litologo-mineralogicheskoye izucheniye glin Maykopskoy svity Prikaspiyskoy neftenosnoy Oblasti Azerbaydzhana.

[2] The Maikop formation is as much as 1,000 m thick in the Cis-Caspian region.

argillaceous fraction, > 0.01, contains quartz, feldspars, clay minerals (hydromica, montmorillonite, kaolinite), a small amount of volcanic glass, glauconite, muscovite, and chlorites; the heavy fraction contains chiefly pyrite (locally up to 95%), muscovite and chlorite, and a group of stable minerals (garnet, tourmaline, rutile). Picotite, titanite, glauconite, amphiboles, and others, also have been noted.

Siderite intercalations are characteristic of the Maikop formation in the central part of the province.

Upper Maikop deposits have been well developed here; in the mineral composition of their terrigenous components, they differ but little from the lower Maikop deposits. They are marked by the preponderance of clay minerals of the montmorillonite group in the light fraction, and a low lime content in the component rocks.

The silicate chemical analysis was performed on the 0.001 mm fraction. The chemical analyses of hydrochloric acid extracts were performed to ascertain the chemical makeup of the Maikop formation as a whole.

The Amirkhanly area clays are marked by a high SiO_2 content (Table 1), 53.4 to 61.6%; Al_2O_3, 19.8 to 25.7%. This increase in Al_2O_3 is apparently connected with the presence of hydromicaceous clay minerals; but the increase in SiO_2 (samples 550, 551, etc.) is due to terrigenous quartz components (chalcedony, opal), whose presence in the clays has also been microscopically confirmed. The TiO_2 content ranges from 0.66 to 87%. The average Fe_2O_3 content in the clays is 2 to 3%. Also typical is the low content of FeO, MnO and CaO. The amount of Na_2O ranges from 0.51 to 2.0%, which is usually characteristic of montmorillonite clays; the increase in K_2O content, in places as much as 4%, is due to the presence of hydromicaceous minerals, as confirmed by microscopic study.

The amount of H_2O in these clays ranges from 4.4 to 6.8%. This somewhat depressed water content is due to contamination with hydromicaceous clays. It is of importance that the SiO_2/R_2O_3 ratio in these clays is four or more, decreasing to 3.9 only in the lower part of the section.

Judging from the results of the silicate chemical analyses (Table 2), the Saadan and Amirkhanly clays differ little from each other, in composition. Their SiO_2/R_2O_3 ratio ranges from three for upper Maikop to 4.6 for the lower.

The amount of Na_2O in the Saadan and Amirkhanly clays is small, 0.1 to 0.8%, while that of K_2O is somewhat high, 2.05 to 3.67%.

The data of the analyses of hydrochloric acid extracts are listed in Table 3.

The upper Maikop (Amirkhanly) is represented by non-calcareous clay, with the CaO content not exceeding 1.5%, but their Fe_2O_3 content is somewhat high, up to 4%, with about 2.0% of SiO_3''. Lower Maikop clays are marked by a high CaO content, 1.8 to 23.6%, and a depressed SO_3'' content, 0.2 to 0.9%.

It has been established, as a result of the recomputation of CaO, MgO and SO_3'' for calcite, magnesite, and gypsum (anhydrite) that the $CaCO_3$ content in upper Maikop clays is very low (up to 0.5%). It increases in lower Maikop beds, beginning with sample No. 515, to reach 23.7% locally and to 42.2% in marly beds. The $MgCO_3$ content for the section rises sharply to 2.55%; while that of gypsum (anhydrite) decreases from top to bottom, from 3.2 to 0.4%.

Hydrochloric acid extracts from the Saadan section clay (Table 4) contains Al_2O_3, 1.74 to 7.6%; Fe_2O_3, 1.42 to 4.5%; CaO, up to 2%. It increases sharply, beginning with sample No. 645 downward (lower division), locally as much as 41.2%. According to the conversion, the amount of $CaCO_3$ in the upper Maikop division is insignificant, not exceeding 2.7%. In the lower division, on the other hand, it increases sharply in individual samples, reaching 70% (sample 636).

The $MgCO_3$ content ranges from 0 to 2.5%. It should be noted that the $CaSO_4 \cdot 2H_2O$ (anhydrite) content increases sharply in the upper part of the section, up to 4.6%, decreasing again to 1.5%, in the lower part.

Analyses of water extracts were carried out to study salt components. It was determined that the HCO_3' content ranged from 0.04 to 0.14% (Amirkhanly) and from 0.01 to 0.6% (Saadan); that of Cl', 0.02 to 0.07% and 0.0 to 0.04%, respectively; SO_4'', 0.07 to 0.05% and 0.15 to 0.36%; Na + K, 0.08 to 0.26% and 0.04 to 0.19% (Table 5).

The amount of C_{org} in the Amirkhanly section ranges from 1.37 to 4.91%, increasing from 2.55 to 5.07% in the Saadan section.

Thermograms of the Amirkhanly [1] clays are generally marked by three endothermic and two exothermic effects (Fig. 1).

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[1] The < 0.001 mm fraction was thermally analyzed.

Table 1

Chemical composition of the < 0.001 mm fraction of Maikop clays from the Amirkhanly region of the Cis-Caspian oil province (in %)

Nos.	Sample Nos.	Well Nos.	Components															
			H_2O^-	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	MgO	MnO	Na_2O	K_2O	SO_3''	H_2O^+	Loss on ignition	Sum	$SiO_2:R_2O_3$
1	509	8	6,10	56,64	0,70	23,10	3,35	0,45	0,08	2,10	0,02	0,51	3,16	0,41	4,85	3,83	99,20	4,0
2	510	8	6,34	60,64	0,73	20,15	2,96	0,88	0,08	2,12	0,03	0,85	2,63	0,68	5,01	2,95	99,71	4,8
3	512	8	5,98	57,90	0,70	21,54	2,92	1,34	0,08	2,76	0,03	0,63	3,68	0,60	4,16	3,98	100,34	4,2
4	514	10	6,36	59,07	0,76	19,92	2,78	0,90	0,16	2,09	0,03	1,09	2,46	0,78	7,01	3,02	100,07	4,8
5	515	10	3,87	54,84	0,80	25,84	2,24	0,81	0,15	0,71	0,03	1,28	1,50	None	7,67	3,63	99,50	3,6
6	522	36	6,22	59,82	0,76	20,43	2,12	1,12	0,08	2,05	0,02	1,04	1,82	0,33	6,41	3,19	99,19	4,7
7	532	36	5,67	56,56	0,81	23,64	1,94	1,34	0,07	2,26	0,03	1,00	1,67	0,25	5,68	4,08	99,33	3,90
8	536	36	6,50	57,88	0,87	22,47	2,08	1,24	0,08	2,42	0,03	1,01	1,87	0,05	5,85	3,65	99,50	4,2
9	538	36	6,80	57,53	0,82	22,45	2,88	0,90	None	2,04	0,02	0,90	2,12	0,31	7,00	2,55	99,52	4,0
10	541	30	4,39	53,44	0,72	25,68	2,65	0,87	None	1,74	0.03	1,40	2,31	0,15	6,05	4,63	99,67	3,5
11	543	30	4,97	55,28	0,66	23,04	2,18	1,32	0,07	1,98	0,02	0,84	2,58	0,30	6,53	4,60	99,40	4,0
12	548	30	5,80	56,45	0,67	22,59	2,71	0,78	0,09	2,01	0,02	1,07	2,45	0,37	5,50	4,91	99,61	4,2
13	550	22	5,47	61,64	0,69	20,39	1,91	1,00	0,04	1,90	0,02	0,52	2,36	0,48	6,02	3,44	100,41	5,1
14	551	22	5,91	61,60	0,67	19,92	2,84	0,89	None	2,56	Traces	1,64	1,65	None	6,43	1,37	99,57	5,0
15	552	13	6,33	59,07	0,73	21,52	3,92	0,90	0,08	2,62	0,02	2,00	1,78	None	6,09	1,27	100,00	4,2
16	553	13	7,14	57,70	0,74	22,99	3,16	1,02	None	2,42	0,03	0,80	3,50	0,29	4,01	3,78	100,54	4,0
17	556	23	5,15	59,16	0,75	19,78	2,74	0,88	0,07	1,90	0,02	0,92	2,36	0,36	6,76	3,98	99,68	4,8
18	559	23	6,38	58,82	0,77	21,69	2,99	0,89	0,16	2,44	0,02	0,54	2,23	0,08	5,65	3,14	99,41	4,2
19	560	24	6,45	56,60	0,75	22,57	4,56	0,45	0,15	2,72	0,03	0,63	3,83	0,08	5,20	2,26	99,83	3,9
20	570	24	6,23	56,17	0,76	23,02	3,10	0,67	None	2,52	0,02	0,61	3,75	0,10	5,40	3,22	99,44	3,9

Note: Comma represents decimal point.

Table 2

Chemical composition of the < 0.001 mm of the Maikop clays from the Saadan region of the Caspian oil province (in %)

Formation	Sample No.	Well No.	Components.															Ratio $SiO_2:R_2O_3$
			H_2O^-	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	CaO	MgO	MnO	Na_2O	K_2O	SO_3''	H_2O^+	Loss on ignition	Sum	
Upper Maikop	608	76	6,93	54,25	0,64	20,88	5,18	1,46	0,04	2,20	0,02	0,10	2,05	2,04	8,03	3,30	100,19	3,0
	614	76	7,14	55,20	0,67	21,19	3,98	1,37	0,04	2,20	0,02	0,22	3,00	2,28	7,0	3,32	100,47	4,0
	616	76	7,64	57,54	0,80	18,95	5,18	1,12	0,05	2,74	0,03	0,68	2,54	1,50	6,22	2,54	99,85	4,6
	675	78	5,66	52,88	0,75	24,66	3,32	1,22	0,05	2,26	traces	0,25	3,67	0,45	5,93	3,83	99,27	3,4
Lower Maikop	674	78	5,50	52,74	0,78	24,45	2,84	1,55	0,14	2,10	traces	0,34	3,20	0,72	6,80	3,91	99,56	3,4
	669	78	5,70	53,97	0,78	22,70	3,03	1,45	0,13	2,10	0,03	0,82	2,80	0,90	6,90	3,85	99,46	3,7
	665	78	6,85	55,88	0,73	22,77	2,10	1,35	0,14	1,97	0,01	0,58	2,96	0,61	6,70	3,70	99,50	4,0
	645	78	5,87	56,30	0,60	21,45	2,66	1,22	0,31	2,27	0,02	0,61	2,89	0,94	7,30	3,32	99,89	4,1

Table 3

The results of analyses of hydrochloric acid extracts from the Maikop clays from the Amirkhanly region (in %)

Component	Sample 510	Sample 511	Sample 515	Sample 527	Sample 531	Sample 532	Sample 536	Sample 538	Sample 543	Sample 545	Sample 550	Sample 551	Sample 553	Sample 558
Al_2O_3	3,30	3,98	3,54	1,16	2,96	4,92	5,61	6,62	6,18	2,12	4,82	6,68	3,02	1,94
Fe_2O_3	4,00	2,08	2,88	1,10	2,04	3,60	3,14	2,38	2,54	3,70	2,52	1,90	2,20	2,66
CaO	0,70	1,20	3,66	9,50	1,82	4,80	2,58	12,64	13,04	13,62	7,52	23,70	23,56	9,88
MgO	0,80	1,28	0,92	0,94	1,24	2,36	1,60	1,01	1,20	1,21	0,50	1,02	0,72	1,58
SO''_3	1,90	1,92	0,68	0,90	0,60	0,50	0,20	0,30	0,44	0,34	0,64	0,20	0,68	0,68
CO_2	1,25	3,42	2,63	6,80	2,40	5,35	4,85	8,40	9,45	10,10	5,96	9,58	12,28	8,54
	6,95	1,21	6,00	4,03	4,85	3,40	4,16	5,10	6,03	9,05	3,66	6,14	5,69	2,73
Insoluble residue	81,44	85,47	79,90	75,31	84,54	75,31	78,45	63,37	61,58	60,13	74,44	52,85	52,00	72,10
Sum	100,34	100,56	100,21	99,74	100,45	100,23	100.59	99,82	100,46	100,27	100,06	100,07	100,15	100,11
H_2O at 100°	0,90	1,53	0,92	1,83	0,10	0,30	0,45	0,10	0,52	1,0	0,1	0,10	0,10	1,43
As computed for $CaCO_3$	None	0,36	5,68	15,76	2,49	7,92	4,34	22,12	23,50	23,83	12,58	31,94	41,04	16,69
$MgCO_3$	1,68	2,68	0,30	0,98	2,50	2,48	1,53	—	—	—	1,05	2,14	—	3,31
$CaSO_4 \cdot 2H_2O$ (anhydrite)	2,15	3,17	1,44	1,93	1,28	1,07	0,42	0,64	0,92	0,71	1,35	0,42	1,44	1,44

Note: Comma represents decimal point.

Table 4

Results of hydrochloric acid extract analyses of the Saadan area Maikop clays (%)

Components	Upper Maikop					Lower Maikop				
	Sample 604	Sample 614	Sample 616	Sample 675	Sample 674	Sample 672	Sample 669	Sample 665	Sample 645	Sample 636
Al_2O_3	5,24	4,98	7,66	4,80	4,96	3,90	3,68	4,82	3,54	1,74
Fe_2O_3	4,44	4,40	4,14	3,18	2,88	2,24	2,42	4,50	3,46	1,42
CaO		0,80	1,54	1,22	4,66	None	None	1,94	12,48	41,16
MgO	0,25	1,02	1,70	0,90	2,02	0,72	0,68	1,08	2,08	0,52
SO''_3	2,00	2,06	3,04	0,54	1,04	0,48	1,06	0,74	0,36	0,70
CO_2	0,30	1,10	1,30	0,95	0,56	0,83	0,49	2,43	10,39	32,64
Loss in heating	6,58	5,10	5,50	6,85	7,13	6,24	6,79	7,54	3,52	—
Insoluble residue	80,59	79,78	75,30	81,25	77,37	85,15	84,56	76,75	64,61	22,08
Sum	99,40	99,23	100,18	99,69	100,62	99,56	99,68	99,20	100,44	100,26
	4,46	2,80	3,79	1,38	1,51	1,78	1,42	1,30	1,27	0,80
Recomp. results $CaCO_3MgCO_3$	None	—	2,72	1,50	1,26	None	None	2,72	21,77	72,37
	0,57	2,11	2,50	1,29	—	1,51	0,96	1,52	2,11	1,50
$CaSO_4 \cdot 2H_2O$ (anhydrite)	None	2,67	4,62	1,16	2,24	—	—	1,59	0,76	1,50

Table 5

Results of water extract analyses of the Maikop clays (%)

Sample	Well No.	Salt component content						By the difference	Sum of salts	Solid residue
		CO_3''	HCO_3'	Cl'	SO_4''	$Ca^{\cdot\cdot}$	$Mg^{\cdot\cdot}$			
The Amirkhanly area										
508	8	None	0,016	0,050	1,126	0,123	0,150	0,129	1,594	1,920
512	8	»	0,038	0,019	0,275	0,033	0,007	0,106	0,478	0,500
514	10	»	0,038	0,075	0,489	0,080	0,012	0,182	0,876	0,974
515	10	»	0,044	0,052	0,600	0,053	0,010	0,257	0,016	0,170
522	36	»	0,055	0,054	0,415	0,028	0,010	0,203	0,765	0,786
532	36	»	0,091	0,030	0,1288	0,030	0,006	0,145	0,544	0,520
548	22	0,001	0,084	0,044	0,274	0,003	0,014	0,163	0,541	0,490
552	13	None	0,143	0,023	0,070	0,003	0,010	0,083	0,266	0,250
553	23	»	0,073	0,024	0,255	0,003	0,007	0,148	0,474	0,500
The Saadan Area										
616	76	None	0,008	0,132	1,054	0,115	0,091	0,288	1,684	1,934
636	78	0,011	0,066	0,038	0,037	0,025	0,003	0,041	0,188	0,266
665	78	None	0,038	0,018	0,399	0,018	0,007	0,182	0,583	0,650
669	78	»	0,010	0,013	0,357	0,010	0,009	0,154	0,548	0,554

Note: Comma represents decimal point.

The first endothermic break occurs in the 0 to 120° C interval and is related to the iberation of hygroscopic water. This effect s fairly intensive compared with the second ndothermic break; this is especially true for ontmorillonite clays (samples 548, 533, 41, 556, etc.).

All thermal curves exhibit a fairly disinct exothermic effect in the 290 to 310° C nterval, connected with combustion of organc matter (including the petroleum bitumens). ome samples (541, 548) register an additional exothermic effect[1] in temperature interval 445 to 465° C, caused by the presence of organic matter (petroleum bitumens) and pyrite.

The second endothermic effect, between 485 and 625° C is related to the OH water in the lattice, and to a partial destruction

[1] The small endothermic break at 335°, in the thermal curve of sample 548 is probably due to the liberation of a residual interlayer water of the zeolite type.

of the latter.

Some of the clays (samples 541, 548, 556) show a sharp exothermic break in the interval between 655 and 605° C, apparently caused by the presence of siderite. It is of interest that the same samples exhibit an endothermic break, between 625 and 740° C, reflecting the liberation of the balance of the OH radical in the clay lattice.

All thermal curves exhibit a double thermal effect of an endothermic break between 840 and 870° C, changing to an exothermic peak between 895 and 930° C.

On the whole, judging from the thermal curves, clays of samples Nos. 541, 548, 556 and others are definitely montmorillonitic, while the curves of samples Nos. 509, 552, 559, etc., reflect a mixture of the montmorillonite and hydromica clay minerals. Thermal curves for the Saadan area upper Maikop clays (Fig. 1) are characterized by the following features.

The first endothermic break, reflecting the loss of the interlayer water, is fairly sharp in all curves, between 50 and 140° C. The second endothermic effect occurs at 585° C, with the maximum within a 565 to 585° range, depending on the sample. The third endothermic break, of low intensity, takes place between 830 and 910° C. It marks the complete dehydration of clay minerals and the destruction of the lattice. Typically, the third endothermic break is accompanied by an exothermic one, between 870 and 960° C, most likely reflecting the crystallization of amorphous products of the decomposition of montmorillonite, in dehydration.

The 310 to 350° C exothermic peak marks the combustion of light bitumens. Thermal curves of samples Nos. 614, 674 and 675 show a complex exothermic effect between 390 and 475° C connected with combustion of organic matter and the presence of pyrite.

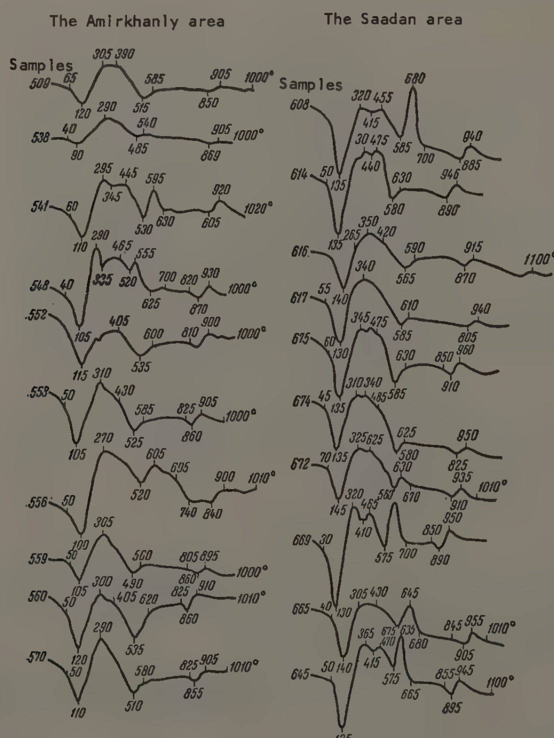

FIGURE 1. Thermograms of the Maikop formation clays (fraction < 0.001) from the Cis-Caspian oil province.

Thermal curves of the lower and upper Maikop differ somewhat in the 575 to 675° C interval where the lower Maikop curve has a sharp exothermic effect of siderite, which is a good correlative marker [9]. The general aspect of the lower Maikop clay thermograms is definitely montmorillonitic.

The study of the loss of weight in heating the Saadan clays, reveals that the dehydration curves are of the same type, for all samples, and have the following characteristics: The maximum loss in weight in dehydration of clays, up to 1,000° C, is 14%. As determined from the dehydration curves (Fig. 2), the loss of weight takes place in these clays at three stages: 1) low-temperature water, up to 6%, is liberated between 50 and 150° C; 2) a continuous loss of weight, chiefly at the expense of the OH radical of the lattice, amounting to 7 to 9%, in the 150 to 700° C temperature interval; this is reflected in an inclined, ascending segment of the curve; 3) complete dehydration of clays takes place in the 700 to 1,000° C interval, where the balance of the lattice water is liberated; this balance is 1 to 2% of the total, and is marked by a nearly horizontal shelf in the curve.

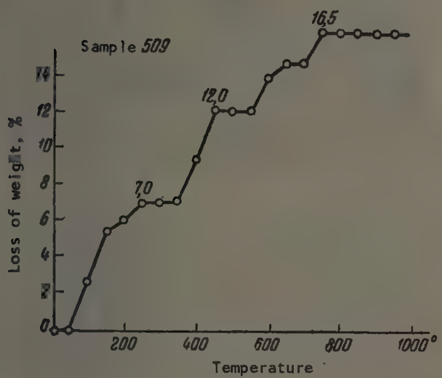

FIGURE 2. Dehydration curve of the Saadan section Maikop clays.

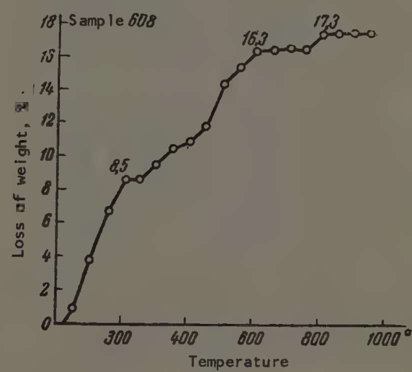

FIGURE 3. Dehydration curve of the Amirkhanly section Maikop clays.

In the Amirkhanly upper Maikop clays, the overall loss of weight in dehydration up to 1,000° C is 16.5%. Judging from the dehydration curves, it also runs in three stages: 1) a continuous loss of weight takes place between 50 and 250° C, chiefly because of the liberation of the interlayer water which accounts for 7%; 2) a sharp loss of weight takes place between 350 and 450° C, because of a 5% water loss; 3) finally, complete dehydration of clays takes place between 550 and 800° C.

Dehydration curves of the lower Maikop clays are characterized by three steps: 1) a 6% loss of weight at the 50 to 200° C interval; 2) a further loss of weight takes place with the liberation of the molecular water, in the 450 to 550° C interval, amounting to 4 to 5% of the total water loss; and 3) dehydration occurs between the 700 and 800° C interval, where the weight loss is 2 to 3% and complete dehydration takes place.

We have succeeded in differentiating the Amirkhanly section upper Maikop clays in the upper and lower divisions by the dye method of N. Ye. Vedeneyeva and M. F. Vikulova, and by the determination of pH and Eh of water suspension of clays. Clay materials of the upper division take on colors from purplish blue to blue when treated with blue methylene (MG), flocculating in small flakes. With added KCl, the suspension turns blue, at times with a pale blue tint, and the precipitate is massive. The spectrum curves of these clays, dyed with MG, usually show a well-defined maximum, between 580 and 600 millimicrons, while the MG + KCl curves are characterized by a maximum at about 670 millimicrons. The The pH and Eh of water solutions of upper Maikop clays show the alkaline-reducing character of the medium.

The lower division clays, with MG and MG + KCl, usually take on purple-blue to blue colors, and give a precipitate of small flakes, rarely in a dense blue mass. Their mineral composition is montmorillonite and

hydromica, with montmorillonites predominating in the upper half. The spectrum curve exhibits, besides the "twin" maximum, a clearly defined one at about 670 millimicrons.

An application of the dye method to the Saadan section clays gave similar results. The upper division clays are blue to light-blue, with MG, and leave a sediment of small flakes. The addition of KCl turns the suspension bluish green, with a gel-like to dense precipitate. Montmorillonite clay minerals predominate, with an admixture of hydromicas. Water suspensions are marked by a high Eh and a depressed pH. The spectrum absorption curves, for clays dyed with MG and MG + KCl, exhibit a maximum at 565 millimicrons. Treated with MG, the lower Maikop clays turn blue to light blue; at times green, with addition of KCl. The precipitate is gel-like to dense.

Two mineral zones have been identified in the section by their clay mineral content. The upper part is montmorillonite and hydromica impurities. The determination of pH and Eh of water solutions show the alkaline-reducing character of the medium.

According to electron-microscopic data, the Amirkhanly area clays (Fig. 4) contain finely-dispersed, compact, flake-like montmorillonite mineral particles, along with distinct semi-transparent scales and sharply-defined tablets of hydromica minerals, and with rare, somewhat modified hexagonal tablets of kaolinite and elongate prismatic grains of halloysite. There are occasional nontransparent well-defined rhombs of carbonates and specks of organic matter.

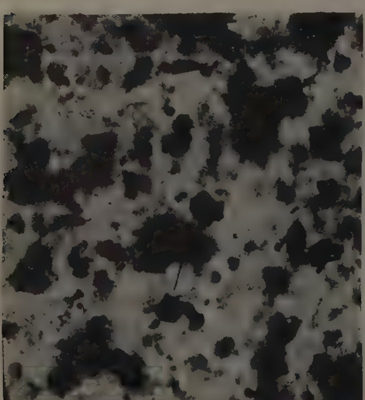

FIGURE 4. Sample No. 551 of the Amirkhanly region clay Electron-microscope; 7,200 X.

The electron-microscopic study of the Saadan area clays (Fig. 5) has revealed coarse flake-like particles, nearly opaque, and their aggregates, caused by the presence of montmorillonite and partly by beidellite (sample No. 608). Well-defined tablets of hydromica were also noted, with rare hexahedrons of kaolinite, their fragments, and

FIGURE 5. Sample No. 363 of the Saadan region clay. Electron microscope; 7,200X.

well-defined rhombs of carbonates. There are typical vague, opaque coarse particles, with sharp, transparent, brush-like growths coming out of them (sample 636). This probably is a phase of transformation of montmorillonite clay minerals to hydromicas.

All samples exhibit numerous small quasi-transparent scales, apparently related to the presence of organic matter. Sample No. 636 shows a distinct long opaque rod, related here to the presence of halloysite.

CONCLUSIONS

1. It has been determined, on the basis of a comprehensive study of clays by various methods, that the principal rock-forming clay minerals of the Maikop formation from the Cis-Caspian oil province are montmorillonite, and to a lesser extent beidellite and hydromica. There are impurities of kaolinite, halloysite and glauconite.

On the whole, the upper division of the Maikop formation is montmorillonitic, with impurities of beidellite and hydromica; the lower division is montmorillonite and hydromica.

2. Among the syngenetic minerals, pyrite, siderite and glauconite have been identified.

3. As a result of dyeing of clays and the determination of pH and Eh, the Maikop section may be subdivided into two, locally three, zones of definite assemblages of clay minerals and of different pH and Eh values of their water suspensions.

4. Oily components predominate in bitumens of the Maikop formation, where they are little or not at all oxidized. The high organic carbon content, locally up to 5%, is characteristic.

5. On the whole, lower Maikop deposits of the Cis-Caspian province are represented by calcareous sedimentary clays. The appearance of hydromica clay minerals, especially in the lower part of this division, suggests a normal marine basin of deposition. The presence of pyrite derived from organic matter, the strong carbonatization of the rocks, and the presence in them of scattered ferruginous carbonates (e.g., siderite), suggest a preponderance of definitely reducing conditions, during the early Maikop time -- namely of hydrogen sulfide and sulfide-siderite geochemical facies. The mineral composition of clays is represented here by montmorillonite and hydromicas.

6. Upper Maikop deposits are represented by a littoral arenaceous argillaceous facies changing to a clay facies, in the southeast. The predominance of montmorillonite clay minerals, the lack of carbonates ($CaCO_3$) in clays, along with the presence of pyrite and organic matter, suggest an alkaline, reducing environment (sulfide geochemical facies).

7. Thus, the Maikop age, especially its first half, had favorable geochemical and lithologic conditions for an adequate accumulation of organic matter and for its subsequent conversion into oil.

On the basis of this study, the Maikop formation is best suited among Azerbaydzhan tertiary rocks to be a probable source of oil.

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Received March 19, 1958

NEW DATA ON THE GEOLOGIC STRUCTURE AND DEVELOPMENT OF THE DONBAS PERIPHERY[1]

by

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A broad belt of deeply buried sediments between the Ukrainian and Voronezh crystalline massifs, from the Pripyat' downwarp in the west to the Yergeni flexure in the east, has been named the Donets downwarp. The history of the development of principal views on its tectonics has been related in detail by N. S. Shatskiy [1].

Recent studies lead to the conclusion that the Donets downwarp is a blind branch of the Crimea-Caucasus Hercynian folded zone wedging in a narrow band into the East European platform. Parallel with the thinning of Carboniferous deposits, the intensity of folding decreases from the inside corner of the platform, in the east, toward the Dnieper-Donets depression and the Chernigov swell, in the west. This decrease in the intensity of folding has also been helped by transverse barriers or bridges in rigid Precambrian folded zones, of a meridional trend, which provide a sturdy frame for the Hercynian-Alpine diastrophism. The Kharkov-Pavlograd bridge splits the Donets downwarp in two: a geosyncline in the east and a platform in the west.

Significant during the initial stages of the geosynclinal downwarping was a migration of provinces with stable marine conditions and of zones of the maximum sedimentation. The stable marine province was a cryptodepression, i. e., a low in the earth's crust under the highest sea level for the time. Chiefly carbonate deposits were laid down within that cryptodepression. Inasmuch as a rhythmic carbonate sedimentation took place in the Donbas, marine cycles must have developed or even predominated within the cryptodepression.

The location of Devonian cryptodepressions apparently coincided with the zone of maximum sedimentation and was associated with the axial part of the Donets downwarp. In Upper Devonian, the near-Azov region of the Ukrainian crystalline massif was marked by intensive volcanism; it appears to have been a part of the Thetis geosynclinal province. A Tournaisian cryptodepression was located in the axial part of the downwarp, at the approximate site of the present Donbas.

In the beginning of the Visean, the marine cryptodepression migrated toward the southern wing of the Donets downwarp, in the area of Yelenovka -- Novo-Troitskoye -- Karabuka villages. Stable marine conditions persisted there till $C_1^V f$ time. The shifting of the deepest sea close to the near-Azov part of the Ukrainian crystalline massif was a harbinger of the ancient Hercynian folding in the Crimea-Caucasus geosynclinal province.

The results of drilling the Peschanokopskaya, Novominskaya, Yeyskaya, Sengilevskaya, and Ipatovskaya boreholes confirmed the presence of an ancient Hercynian Crimea-Caucasus geosynclinal province, south of the Donets geosynclinal downwarp, and separated from it by the Near-Azov massif.

This geosynclinal province underwent orogenic movements, during a pre-Upper Visean phase. After the ancient Hercynian folding, two foredeeps began to form there, the Donets in the west and the Ural-Emba in the east -- symmetrically in relation to the interior corner of the platform. As a result of the ancient Hercynian diastrophism, the marine cryptodepression shifted, in the Upper Visean, over the southeastern slope of the Voronezh massif. As shown by A. Z. Shirokov [6], beginning with the Upper Visean and up to the close of the Carboniferous, the maximum downwarping was concentrated in the main Donbas anticline zone. The latter, along with a series of structures in its northwestern extension, originated during Araucarites time [7].

As pointed out before, the maximum

[1] Novyye dannyye o geologicheskom stroyenii i istorii razvitiya okrain Donetskogo basseyna.

sedimentation zone migrated north of the main anticline, in the beginning of Upper Permian, to coincide with a line through the northern Anticline -- Petrovsko-Slavyansk anticline. The development of the geosynclinal part of the Donets downwarp was quite different from that of typical foredeeps. The axes of the latter tended to migrate gradually toward the platform, which precluded the concentration of sediments in narrow linear zones.

An effect of the younger Hercynian diastrophism was formation of the southern, main, and northern anticlines, in the deepest reaches of the Donets downwarp; formation of the middle massif, along its southern fringe; formation of the northern and southern Donbas foredeeps, to the north and southwest, respectively; and formation of the Dnieper-Donets depression, to the west (taphrosubgeosyncline of V.G. Bondarchuk). [Tr. note: Taphro-structures are associated with deep-seated major faulting]. It is possible that east of Yergeni, the Donets downwarp is separated by a central massif from a bay-like branch of the ancient Hercynian geosynclinal province.

The apex of the Voronezh crystalline massif is faulted off in the south, along the Novyy Oskol-Boguchar line (the A.A. Dubyanskiy fault). South of the fault, is the downthrown side of the Voronezh shield, where the crystalline basement plunges from +43 m (Rossosh') to -265 m (Kantemirovka), which is 4.7 m to a kilometer. Here, the Precambrian massif is overlain by a Lower Carboniferous argillaceous-carbonate sequence (300 to 400 m thick) and by Mesozoic and Cenozoic deposits (up to 140 m thick). The next bench to the south stretches along the Valuyki-Kamenka-Chertkovo-Napolovskaya line. This is the edge of the Voronezh shield, with evidence of folding -- such as the Belokurakinsk structural terrace, Kamensk dislocations, Napolovsk anticline, etc., -- present along it. East of the Napolovsk anticline, the trend of the bench changes from latitudinal to meridional (Fig. 1). Dislocations are present here, too: over ten small uplifts, from Veshenskaya station to Uryupinsk. A wedge of a C_2^1 - C_2^2 Carboniferous sequence (50 to 60 m thick) is associated with this bench.

The Voronezh shield edge bench adjoins a stable block in the south, or sub-platform, 30 to 60 km wide. The crystalline basement plunges here from 600 to 2,000 m.

The southern edge of that sub-platform is marked by another bench, adjoining in the west the Romny fault zone of the Dnieper-Donets depression. This is one of the ancient benches of the Donets downwarp, formed in the Devonian. Exploratory drilling in the Kupyansk area has established the presence of C_2^7 and C_3^1 Middle and Upper Carboniferous deposits, in the upper side of the bench, 500 to 520 m deep, below Middle Jurassic beds. Triassic, Permian, and Devonian rocks wedge out in the lower part of the bench, where a considerable thickening of Upper Carboniferous sediments also occurs. Adjacent salt diapirs may be present in the lower part of the sub-platform edge bench.

Geophysical study has revealed a transition zone, 40 km northeast of Lugansk, at Verkhniy Menchikov village, where Carboniferous carbonate deposits change to arenaceous argillaceous deposits, with an abrupt thickening from north to south [3].

In the Gorodishchenskaya anticline, 18 km north of Verkhniy Menchikov, Precambrian crystallines have been reached at about 1,400 m. Consequently, the Verkhniy Menchikov bench is a sub-platform edge bench. It runs east, toward stanitsa Perelazovskaya, then veers sharply to the north.

A mobile block wedges in, in a narrow band, in the Millerovo swell area; it merges with the Don-Medveditsa dislocation area to the east. The western chain of the Don-Medveditsa structures stretches as far as the Millerovo swell and terminates in the Verkhne-Tarasovskaya anticline.

Adjacent to the sub-platform and mobile block in the south is the northern Donbas foredeep, whose southern boundary is the northern Donets overthrust and the northern Donbas zone of low domal folding.

The western part of the foredeep is represented by a group of compensatory synclines: the Zmiyev-Balakleya, Borovaya, Kaban'yeva-Slavyanoserbsk, and Chirskaya-Morozovsk (Fig. 1). The formation of these synclines was related to considerable extent to the outflow of Devonian salt; accordingly, they are called here, compensational.

The deepest foredeep synclines are the Zmiyev-Balakleya and Chirskaya-Morozovsk. They carry Permian deposits in their near-axial zone, with gypsum and anhydrite appearing in the latter, east of the Kamensk meridian. The near-axial part of the Morozovsk syncline has been complicated by the Astakhovsk-Skosyrsk swell-like uplift.

In the light of these data, the northern wing of the Donets downwarp consists of a series of sunken, step-like blocks, band-like in plan. Thus, we arrive at the necessity of dividing the subject area into shields and blocks of different types, depending on their mobility.

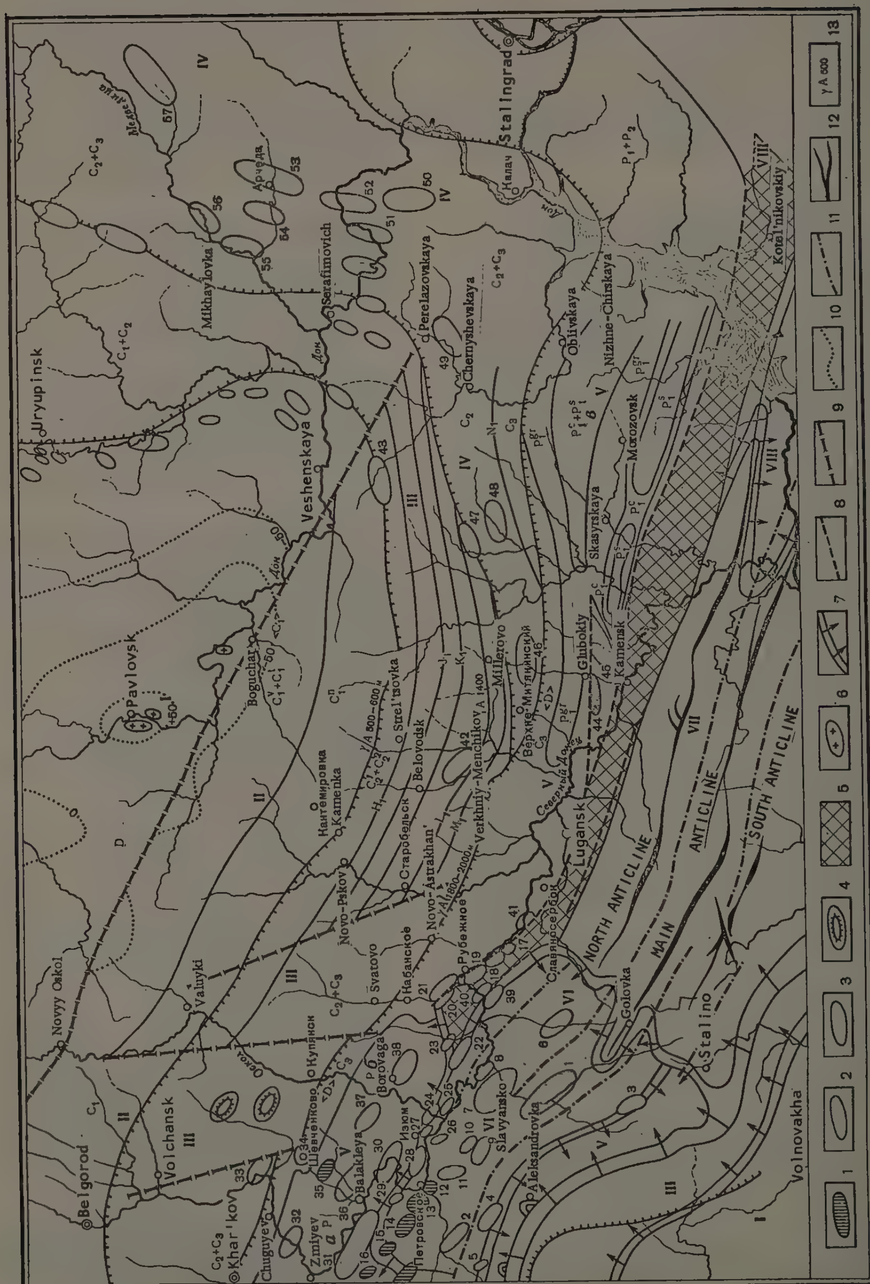
Stalingrad
Uryupinsk
Serafimovich
Veshenskaya
Mikhaylovka
Perelazovskaya
Chernyshevskaya
Oblivskaya
Nizhne-Chirskaya
Kotel'nikovskiy
Morozovsk
Skasyrskaya
Pavlovsk
Boguchar
Millerovo
Glubokiy
Kamensk
Lugansk
Belgorod
Khar'kov
Volchansk
Novyy Oskol
Valuyki
Chuguyev
Zmiyev
Balakleya
Borovaga
Svatovo
Novo-Pskov
Kamenka
Belovodsk
Strel'tsovka
Verkhniy-Menchikov
Novo-Astrakhan'
Slavyansk
Aleksandrovka
Golovka
Stalino
Volnovakha
NORTH ANTICLINE
MAIN ANTICLINE
SOUTH ANTICLINE

FIGURE 1. Index map of the Donets basin.

1 -- salt domes; 2 -- alleged salt structures; 3 -- structures undisturbed by salt; 4 -- Paleogene disturbances; 5 -- low-domal folding zone of the Northern Donbas; 6 -- exposures of crystalline rocks; 7 -- monoclines (benches and flexures; 8 -- thrusts; 9 -- normal faults; 10 -- contours on Precambrian basement; 11 -- synclinal axes; 12 -- linear Donbas folds; 13 -- Precambrian elevations; $C_1^v + C_1^t$ etc., age of rocks underneath the Mesozoic and Cenozoic.

Synclines: a) Zmiyev-Balakleya; b) Kaban'yevsk-Slavyanoserbsk; c) Chirskaya Morozovsk.

I -- the most uplifted areas of the crystalline massif; II -- slope of the Voronezh crystalline massif; III -- stable shields (subplatforms); IV -- mobile shields; V -- Donbas foredeeps; VI -- area of the northwestern plunge of the Donbas (suborogeny); VII — Donets folded system (orogeny); VIII -- easterly extension of the Donets folded system.

Local structures: A. The Main Anticline zone: 1 -- Druzhkovsk-Konstantinovskaya; 2 -- Novomechebilovskaya; B. Southern Anticline zone: 3 -- Ochertinskaya; 4 -- Gavrilovskaya; 5 -- Rudayevskaya; C. Northern Anticline zone: 6 -- Bakhmutskaya; 7 -- Slavyanskaya; 8 -- Nizhnetoretskaya; 9 -- Korul'skaya; 10 -- Novoselovskaya; 11 -- Kamyshevskhskaya; 12 -- Nizhnebereksкaya; 13 -- Petrovskaya; 14 -- Volvenkovskaya; 15 -- Lozoven'kovskaya; 16 -- Alekseyevskaya; D. Zone of Lisichansk-Vol'venkovskaya anticlinal flexure and Middle Anticline: 17 -- Matrosskaya; 18 -- Lisichanskaya; 19 -- Tomashevskaya; 20 -- Kremenskaya; 21 -- Krasnopopovskaya; 22 -- Torsko-Shandrigolovskaya; 23 -- Ternovskaya; 24 -- Krasnooskol'skaya; 25 -- Svyatogorskaya; 26 -- Sukhokamenskaya; 27 -- Izyumskaya; 28 -- Spevakovskaya; 29 -- Protopopovskaya; E. Other structures of the Donets downwarp: 30 -- Chernovodonetskaya; 31 -- Shebelinskaya; 32 -- Mokhnachevskaya; 33 -- Pisarevskaya; 34 -- Shevchenkovskaya; 35 -- Brigadirovskaya; 36 -- Balakleyskaya; 37 -- Severo-Golubovskaya; 38 -- Chernechenskaya; 39 -- Dronovskaya; 40 -- Novodubrovskaya; 41 -- Toshkovskaya; 42 -- Gorodishchenskaya; 43 -- Napolovskaya; 44 -- Astakhovskaya; 45 -- Gusevskaya. F. Don-Medveditsa swell zone: 46 -- Verkhnetarasovskaya; 47 -- Pervomayskaya; 48 -- Adamovskaya; 49 -- Bakachevskaya; 50 -- Sirotinskaya; 51 -- Verkhovskaya; 52 -- Saushinskaya; 53 -- Archelinskaya; 54 -- Vetyutnevskaya; 55 -- Abramovskaya; 56 -- Mironychevskaya; 57 -- Korobkovskaya.

Drilling, geophysics, and subsurface study in the northern limb of the Donets downwarp have revealed the transverse Livensk-Kupyansk, Pavlovsk-Strel'tsovka, Millerovo, Archedinsk-Kotel'nikovskiy and other swells. The most conspicuous is the Pavlovsk-Strel'tsovka swell, adjoining the main Donets folding in the south. These transverse swells, alternating with synclinal troughs, lend a festoon-like aspect to the structure of the entire northern limb of the Donets downwarp. The axes of these transverse folds are 25 to 40 km apart; the fold amplitude, on the Carboniferous, is 60 to 120 m and over.

Sub-latitudinally trending normal faults are suggested by the Paleogene structure in the western part of the area.

Paleogeographic conditions in the northern part of the foredeep were favorable for the formation of reef massifs. It is possible that the surface of the $C_1 - C_2^2$ calcareous sequence may turn out to be broken into many shelves which are commonly associated with oil deposits.

Small deposits of Devonian salt appear along the edge shelf zone; they are sizable in volume only in the northern Donets thrust zone and in the Lisichansk-Volvenkovskaya anticlinal flexure, possibly originating in an ancient fault.

The structure of each compensatory syncline on the northern Donbas foredeep comprises three zones: 1) inner, or southern, adjoining the northern Donets thrust; 2) central; and 3) outer (northern), contacting the zone of step faulting.

Under static pressure, the main body of Devonian salt structures migrated from the compensatory synclines south to the zone of the northern Donets thrust and the Lisichansk-Volvenkovsk anticlinal flexure. A gradual deterioration of the migration conditions for the salt determined the formation of a chain of flat structures of the near-Donets Cretaceous swell, shifted to the north. Permian salt was subsequently deposited along the northwestern periphery of the Donbas and in the southeastern part of the Dnieper-Donets depression. For this reason, two-storied salt structures may exist in the sub-orogenic area, with the Svyatogorsk and Sukhokamensk structures tentatively placed among them.

Seismic work and shallow drilling in the central zones of the Zmiyev-Balakleya and Borovaya synclines have revealed gentle swell-like uplifts: the Mokhnachevskoye, Golubovskoye, and Chernechenskoye. Similar uplifts may be present in the axial zone of the Kaban'yevsk-Slavyanoserbsk syncline

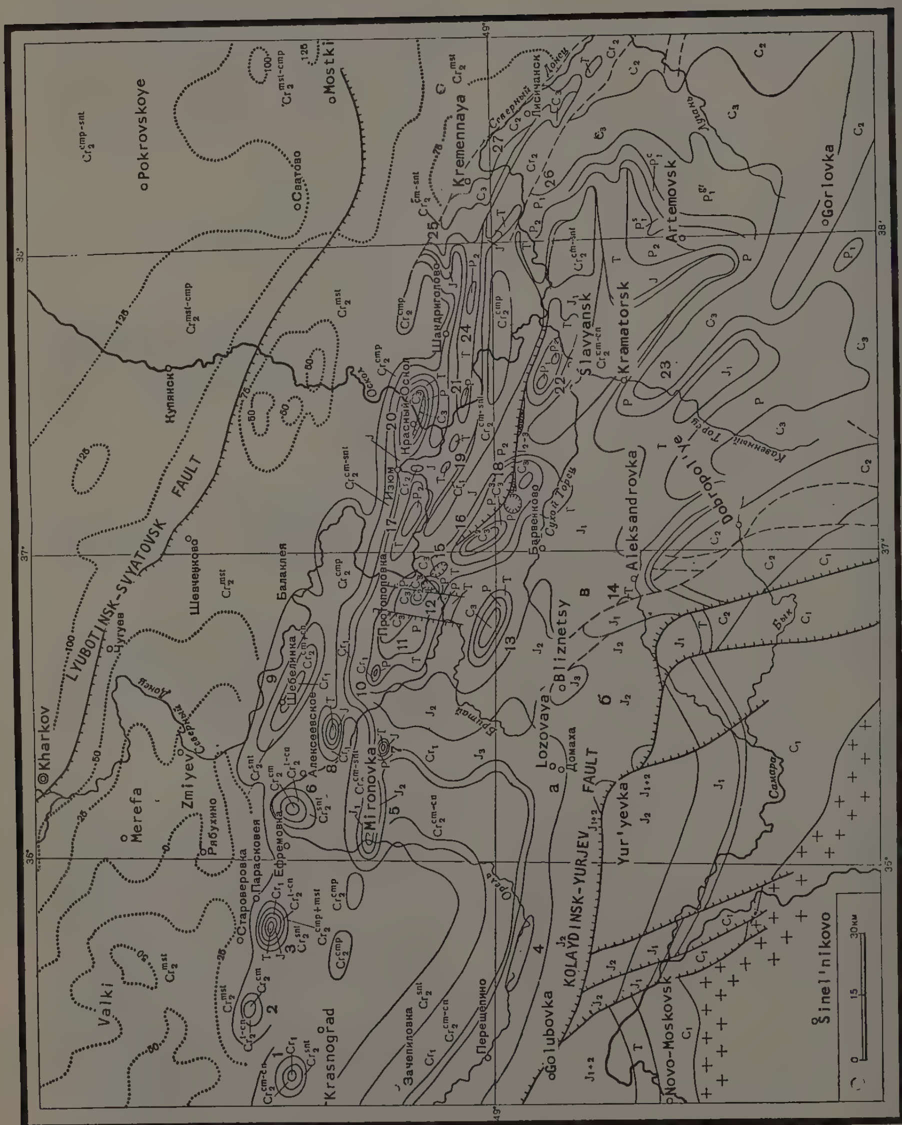
Kharkov
Merefa
Valki
Zmiyev
Krasnograd
Pokrovskoye
Mostki
LYUBOTINSK-SVYATOVSK FAULT
KOLAYDINSK-YURJEV FAULT
Kremennaya
Slavyansk
Kramatorsk
Artemovsk
Gorlovka
Aleksandrovka
Dobropol'ye
Bliznetsy
Lozovaya
Yur'yevka
Golubovka
Mironovka
Novo-Moskovsk
Sinel'nikovo

FIGURE 2. Tectonic map of the western reaches of the Donets Basin

(Compiled by N. F. Balukhovskiy from the data of N. F. Balukhovskiy, V. F. Bliznyuk, B. S. Vorob'yev, B. S. Kovalev, A. K. Kurilik, V. P. Litvinov, A. G. Palamarchuk, L. S. Palets, B. P. Sterlin, V. K. Stovpovoy, L. M. Yarchenko and others).

1 -- Precambrian; 2 -- grabens; 3 -- faults; 4 -- contours on Upper Cretaceous surface; 5 -- breccia; 6 -- overthrusts.

Structures: 1 -- Verkhnelanovskaya; 2 -- Krestischenskaya; 3 -- Medvedovskaya; 4 -- Dubrovogryadinskaya; 5 -- Mironovsk; 6 -- Yefremovsk; 7 -- Belyayevsk; 8 -- Alekseyevsk; 9 -- Shebelinsk; 10 -- Lozoven'kovskaya; 11 -- Volvenkovskaya; 12 -- Petrovskaya; 13 -- Novomechebilovskaya; 14 -- Novobakhmet'yevskaya; 15 -- Nizhnebereksksaya; 16 -- Kamyshevakhskaya; 17 -- Spevakovskaya; 18 -- Korul'skaya; 19 -- Sukhokamenskaya; 20 -- Krasnooskol'skaya; 21 -- Svyatogorskaya; 22 -- Slavyanskaya; 23 -- Druzhkovsk-Konstantinovskaya; 24 -- Torsk-Shandrigolovskaya; 26 -- Dronovskaya; 27 -- Lisichansk-Kremenskaya;

Synclines: a -- Domakhskaya; b -- Lukashevskaya; c -- Yelizavetovskaya.

(Fig. 1). Generally speaking, near-axial parts of major synclines, downwarps, and syneclises have a tendency for upwarping.

The structure of outer (northern) zones of the compensatory synclines is not clear, as yet. It is possible that a small residual body of salt sediments in places was squeezed out of the axial zone, to the north, where it formed, (e.g., at Shevchenkovo, Svatovo), flat primary domes, leaned against the edge bench zone.

The festoon structure of the northern limb of the Donets downwarp, increasingly step-like with depth, the possibility of salt structures in edge zones of compensatory synclines, and swell-like uplifts in their near-axial parts -- are the tectonic peculiarities of the subject area.

In the Devonian and Carboniferous, the northern periphery of the Donbas underwent a struggle between land and sea. The strongest facies changes and thinning of deposits, as well as transgression of the sea, all took place along the benches. Such conditions are favorable for the formation of structural and stratigraphic oil and gas traps.

Lower and Middle Carboniferous coal measures were deposited over a delta plain, in short and rhythmic transgressive and regressive cycles, which is now reflected in the alternation of bottomset and topset delta plains. With the bottomset deposits predominating in the section, the coal content dropped sharply, and barren foreset beds were deposited. In the periods of transgression, the productive coal measure belt shifted into the Dnieper-Donets depression, while chiefly barren or lean coal beds were deposited in the Donets basin. Such are C_1^{V2}, $C_1^4 - C_1^6$, $C_2^1 - C_2^2$, C_2^7, and partly C_2^4. Deltaic conditions favorable for the accumulation of coal prevailed in the Donbas and its western reaches during

$$C_2^3 \text{ and } C_2^5 - C_2^6.$$

A typical feature of the bench-forming process in the northern limb of the Donbas downwarp is the prolonged manifestation of inherited movements, and a gradual decrease in the magnitude of the rock displacement, from older to younger stratigraphic sequences.

In the Middle Devonian, faults originated at the site of modern benches, with the maximum downward movement of individual blocks. Subsequently, the movement along fault surfaces slowed down. Under the conditions of intensive sedimentation, a scarp originated at the fault. It widened upward, to 6 or 7 km, flattening gradually to 1 or 2%. Paleogene and Cretaceous dislocations commonly took place along either side of such a flat scarp.

The northwestern reaches of the Donbas and the northern part of the Old Donbas, taken as a whole, represent a single salt-dome province with a diversity of salt structures. This view was first advanced by A.A. Bogdanov [2].

The geologic history of this region seems to have been as follows. As a result of static pressure, the main and northern Donbas anticlines were formed at different tectonic stages at the close of the Carboniferous and in the Permian. The following structures originated in an extension of the main anticline zone, at that time: Druzhkovsko-Konstantinovskaya, Novomechebilovskaya, Belyayevskaya, Mironovskaya, Pavlovskaya, Sosnovskaya, Verkhnelanovskaya and Yelizavetovskaya (Fig. 2).

The northern anticline zone includes the following structures: Slavyanskaya, Korul'skaya, Petrovskaya, Volvenkovskaya, Lozoven'kovskaya, Alekseyevskaya, Yefremovskaya, Paraskoveyskaya, Medvedovskaya, and Krestishchenskaya. In the northwestern extension of the main and northern anticlines, linear elongated swell-like uplifts arose, each consisting of a chain of anticlines. Salt migrated rapidly from adjacent synclines and depressions and into the crests of newly-formed uplifts. The initial stage of this migration was prompted by the static load, alone.

Subsequently, the salt accumulations over the uplifts were subjected to the inherited upward (magmatic) pressure which reversed the salt movement in the direction of the adjacent synclines and periclinally. Locally, the migration of salt was accompanied by tangential stresses. The combined effect of vertical and tangential forces led to the formation of breccia. The salt domes were complicated by normal faults, overthrust folds, and overhangs.

Four principal stages have been recognized in the development of salt structures: 1) the deposition of an evaporite formation; 2) initiation and development of endogenous folded forms; 3) migration of salt, under static pressure, toward the crests of endogenous structures and toward deep faults; 4) the development of complex salt structures as an effect of the vertical magmatic pressure and the resulting stresses. It follows from this enumeration that salt stocks and deposits must be underlain by buried anticlines.

The vigorous argument as to a possible extension of the Donets folding to the Dnieper-Donets depression -- an argument which antedated the Great Patriotic War -- has finally been brought to a compromise.

A.D. Arkhangel'skiy [1] believed that the Donets folding penetrated deeply into the Dnieper-Donets depression. L.F. Lungersgauzen [4], too, traced the individual Donbas anticlinal zones as far as the depression. On the other hand, M.M. Tetyayev insisted that the linear Donbas folding gives place, in the northwest, to a domal folding and then dies down. Now we know that salt structures of the northwestern Donbas and the southeastern Dnieper-Donets depression are grouped in lines adjoining the principal old Donbas structures in the east -- the main and northern anticlines. The change in folding and its easterly intensification are related to the thickening of Carboniferous deposits and to the plunge of the crystalline basement.

The results of shallow exploratory drilling, carried out recently in the Torsk-Shandrigolovskaya area (under the direction of V.N. Butenko and A.A. Sovinskaya) lead to the conclusion that the Torsk-Shandrigolovskaya fold did not exist in the Upper Permian but was formed during a pre-Triassic (Pfalz) phase. The main folding here was contemporaneous with the development of the main anticline, with the Donets (pre-Bajocian), Novo-Kimmerigian, and Laramide as its principal phases.

The Lisichansk-Volvenkovskaya anticlinal flexure and the northern Donets thrust zone were initiated simultaneously with the main Donbas anticline, and belong to a pre-Permian (Uralian) phase. The growth and development of this folded zone was accompanied here by a gradual southerly shift of the Torsk-Shandrigolovskaya anticlinal crest.

The initiation of the Petrovsk-Slavyanskaya and northern anticlines took place in the Saale (pre-Upper Permian) phase. The vertical stresses in the crest of the northern Donbas anticline squeezed the Devonian evaporite deposits out and northward, where the salt produced a belt of domal folding.

Further study of the history of the Shebelinskaya zone is essential. In our interpretation, this zone is limited by the distribution of the productive Lower Permian salt-bearing formation. The central part of the zone comprises a number of structures (from west to east): Ryabukhinskaya, Shebelinskaya, Chervonodonetskaya, Spevakovskaya (Peresechnenskaya), Svyatogorskaya, Sukhokamenskaya, Nizhnetoretskaya and Bakhmutskaya.

The Shebelinskaya zone and its immediate vicinity is marked by a different depth of structures, with the Ryabukhinskaya anticline the most deeply buried. This anticline heads a peculiar group of structures in and about the Shebelinsk zone (Fig. 3).

Tectonic level	Name of structure and its analogues	Tentative elevation on Carboniferous surface
1	Chervonodonetskaya (Ryabukhinskaya)	−2250
2	Shebelinskaya	−1950
3	Spevakovskaya (Peresechnenskaya)	−1200
4	Protopopovskaya	−1400
5	Svyatogorskaya (Sukhokamenskaya and Nizhnetoretskaya)	−1800
6	Slavyanskaya (Novoselkovskaya)	−1800
7	Volvenkovskaya (Korul'skaya and Novomechebilovskaya)	+120 (formation C_3^3)
8	Petrovskaya	+120 (formation C_2^5)

In recent years, salt, supposedly Devonian in age, has been discovered by drilling along the Donbas periphery. Evaporite beds were found in the crests of the Alekseyevskaya and Berekskaya anticlines.

The Alekseyevskaya anticline, 10 km southwest of district center Alekseyevskoye, the Khar'kov oblast', is a cryptodiapir fold. Breccia of assorted rocks, including salt, has been found below Permian (?) deposits, at 1060 m, and traced to 1,950 m.

The Berekskaya anticline is located between the Kamyshevakhskiy and Petrovskiy domes, in the flood plain of Bereka River. A large gravity anomaly was found here by V.T. Komarevskiy. The author determined the presence of an anticlinal fold here, in 1940. Its crest, made up of Triassic and Jurassic formations, has been complicated by a secondary syncline.

According to V.F. Bliznyuk, this secondary syncline involves the following section: Anthropogene, 20 m; Neogene, 192 m; Kharkov stage, 70 m; Kiev stage, 26 m; Buchak-Kanev deposits, 152 m. Breccia was found below the Neogene, at 460 to 500 m, with salt below 500 or 513 m. The syncline is 7 km long and about 3 km wide (Fig. 4). Superimposed synclines also complicate the crests of the Belyayevskaya and Medvedovskaya structures.

Along the Donbas periphery, salt structures are inferred from indirect evidence, such as the intensity of the inherited movements, expressed in relief, the salinity of the ground water, and the peculiar tectonic structure.

Of interest is the classification of the Donbas diapirs, depending on the stratigraphic level of salt intrusions. Three intrusion levels of the Devonian salt have been recognized: 1) Namurian (Krasnooskol'skaya Petrovskaya, Novomechebilovskaya, Kamyshevakhskaya, Korul'skaya, Slavyanskaya, Druzhkovsko-Konstantinovskaya, Kremenskaya, etc); 2) sub-Upper Permian (Alekseyevskaya, Belyayevskaya, Yefremovskaya, Mironovskaya, and Medvedovskaya); 3) Middle Carboniferous (Lozoven'kovskaya); and 4) Paleogene (Orlovskaya, Nizhneberekskaya, and Gusevskaya).

The presence of beds representing Namurian interval has been inferred from salt in the lower Carboniferous, in the Krasnooskol'skaya and Petrovskaya structures; other stratigraphic levels -- from geophysics and drilling. Cryptodiapirs with a Namurian salt intrusion predominate along the northwestern periphery of the Donbas or Donbas disturbance. At the same time, younger cryptodiapirs with a Paleogene unit occur here. Individual regions of the Donets disturbance are characterized by the absence of beds representing the interval between the Namurian and Paleogene, which suggests the sudden onset of the latter.

As a rule, salt intrusions are bounded by quasi-impenetrable competent rocks. The spasmodic development of salt structures has been observed; the gradual inflow of salt, which causes a smooth movement of rocks, leads to abrupt breaks through incompetent beds. It is quite possible that gas discharges took part in this discontinuous development of salt structures. Metamorphic processes in rocks may cause a systematic rise in gas pressure to that above hydrostatic, and even above rock pressure (Shebelinka). Under such conditions, seismic phenomena or geochemical factors may disturb the equilibrium and cause gas eruptions.

V.F. Bliznyuk believes that cap craters of the Petrovskoye and Nizhnebereksk domes have been formed because of the leaching of salt.

Where a breccia zone is not obviously related to salt tectonics, as in the Korul'sk dome, it should be regarded, according to A.K. Kurilika and Ye. B. Chutko, as

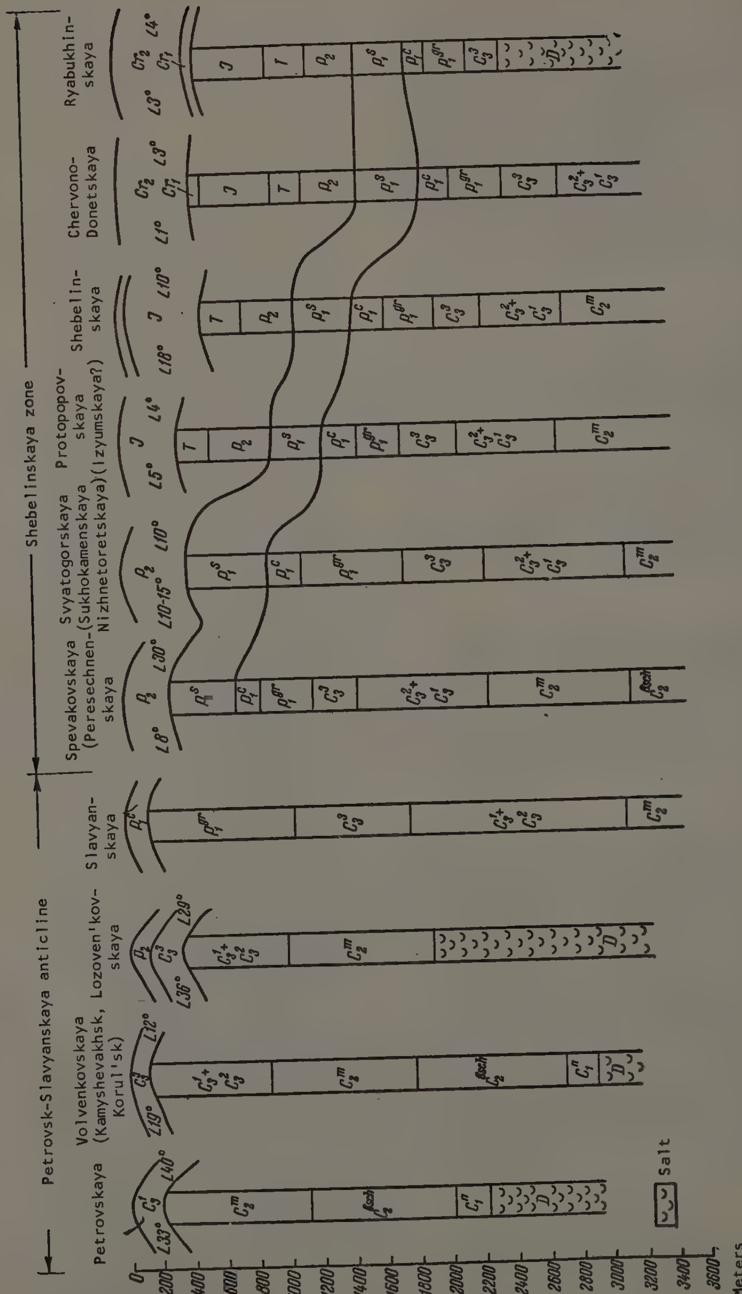

FIGURE 3. Correlation of the peripheral Donbas structures.

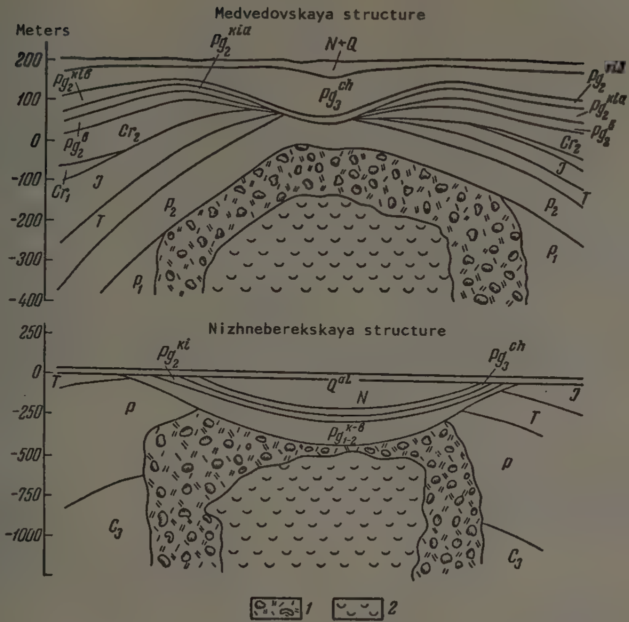

FIGURE 4. Structure of superimposed synclines (after V. F. Bliznyuk and Kh. R. Vidomenko)

1 -- breccia; 2 -- salt.

colluvium of a neighboring piercement dome. The authors do not specify the position of such a dome, since there are no piercement salt structures in the vicinity of the Korul'skoy dome.

Those investigators who deny the participation of gas eruptions in the formation of the Donbas peripheral structures are unable to explain the following facts:

1. The presence of fault zones, filled with breccia and not related directly to salt tectonics (Slavyanskoy and Korul'skoy domes, the Glubokaya River agglomerate area, etc.). The throw of breccia-filled fault zones attains 1,600 m (Petrovskoye dome). On the Korul'skiy dome, the breccia zone is located in a synclinal fold; on the Slavyanskiy dome, it passes from the flanks to the syncline.

2. Hydrothermal processes, associated with faults (Petrovskoye dome).

3. Maastrichtian and Paleocene marine deposits, fringing the graben-like craters, carry very rough fragments of Paleozoic rocks (Glubokaya River agglomerate area).

4. If the graben-like craters have been formed as a result of leaching of salt, why are there no similar forms developed on the Isachkovaya and Romny structures, where the conditions for deep erosion of salt deposits by a Paleogene sea were very favorable?

Subsequent migration of salt along the fault planes is very probable. Its leaching is also possible; however, gas eruptions are the primary factors, in certain areas (Petrovskiy and Slavyanskiy domes and the Glubokaya River agglomerate area).

We interpret the fault zone breccias, trains of erratic rocks about salt structures, the stable associations of Namurian and Paleogene salt units, graben-like craters, and other phenomena, as a peculiar evidence of "sedimentary gas volcanism" in salt dome provinces of fairly consolidated rocks. Rock salt, water, gas and oil form a discrete dynamic system, with the fluids moving in advance of salt and corroding the incompetent rocks, so that the migration of salt is unobstructed.

We shall turn now to the southern Donbas foredeep.

On the basis of the 1950 work of the geophysicist V.S. Zavistovskiy, we postulated an anticlinal flexure in the axial part of the Kal'mius-Toretsk depression.

In 1954, Yu. M. Romanov and G.P. Gutarev shot the Ocheretinskaya anticline having about 1,000 m closure and dips of 11 to 15° on the northeastern limb and 5 to 7° on the southwestern. Thus, it has been confirmed that the southern Donbas anticline continues west, as the Ocheretinskaya, Gavrilovskaya, and Rudayevskaya structures and the Pereshchepino-Lozovskiy swell. The structures are separated here by deep Jurassic synclines. According to I.M. Yamnichenko, Jurassic deposits are up to 650 m thick in the southern foredeep. Upper Cretaceous deposits are uncommon and thin.

In the Upper Cretaceous, the maximum sedimentation zone shifted to the northern Donbas foredeep, where it was as much as 750 m thick (Borovskaya syncline). This is the manifestation of wavy oscillations associated with the development of the Crimean geosyncline.

There are three benches in the southwestern limb of the Donets downwarp. The first one, the nearest to the Ukrainian crystalline massif, passes along line Pavlograd-Volnovakha. The next to the north may be traced along the Kolaydintsy-Yur'yevka. According to L.M. Levenshtein, the Kolaydintsy-Yur'yevka bench adjoins the Krivorozhsk-Pavlovskiy normal fault. An abrupt thickening of Carboniferous deposits and wedging-out of the Devonian are associated with the Kolaydinsk-Yur'yevka bench zone. The third bench of the Donets downwarp southern limb passes through Isachki-Pereshchepino-Samoylovka, to merge with the southwestern limb of the Kal'mius-Toretskaya trough. Three structural noses have been noted in the lower side of this bench: the Samoylovskiy, Grigorovskiy and Novobakhmet'yevskiy, with compensatory synclines between them: (from west to east) Domakhskaya, Lukashevskaya, Yelizavetovskaya, etc. (Fig. 2). The compensatory synclines area is bound on the north and northeast by the southern Anticline zone; in the south, by the Yur'yevka-Krivorozhsk-Pavlovsk bench. This region has been named the southern Donbas foredeep. It is 20 to 45 km wide.

On the whole, the above data give credence to N.S. Shatskiy's concept that the Great Donbas is a typical transversal downwarp, most like the Wichita system, U.S.A., in its tectonic nature and position on the platform. Isolated projections of geosynclinal provinces of the first order (orthogeosynclines) form wedges in the platform. According to Shatskiy, transverse marginal systems or downwarps are closely related to interior angles of platforms. One can concur with Shatskiy in assuming an interior angle of the Russian platform "somewhere in the lower course of the Volga"[5].

Areas the nearest to the interior angle of the Donets downwarp have gone through a cycle of geosynclinal development. This has been reflected in the well-defined oscillatory movements and in the accumulation of thick sedimentary sequences.

Two early Hercynian downwarps, both adjoining the interior angle of the platform, should be recognized in the Hercynian structure of the south part of the U.S.S.R.: the intra-platform Donets, in the west, and the Cis-Caspian in the east. Their geologic development proceeded along different lines.

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Received October 20, 1958

RUBY SPINEL OF THE PEREVAL DEPOSIT, AND ITS SECONDARY ALTERATIONS[1]

by

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Ruby spinel occurs along with blue spinel, in strongly metamorphosed rocks of the southern Cis-Baikal region. It was noted by B.Z. Kolenko [6] in calcifers of the "White quarry" near Ponomarevka siding of the Eastern Siberian Railroad; by N. Voskoboynikova [2] and D.S. Korzhinskiy [7], in the Sludyanka lazurite deposit; by L.M. Lebedev and N.G. Sumin [8] in a block of coarse-grained marble, in the middle course of Slyudyanka River; and finally by P.V. Kalinin [5], in the Smirnov mine near the town of Slyudyanka. The author has studied ruby spinel from the Pereval (Slyudyanka district) marble deposit in great detail, and has succeeded in determining the properties of this mineral and of its alteration products, which is an important step toward an understanding of the overall picture of metamorphic evolution of spinel-bearing marble.

The Pereval cement marble deposit is located on the Slyudyanka-Pokhabikha watershed, seven kilometers from Slyudyanka. It consists of beds and lenses of Archaean carbonate rocks, with calcite, calcite-dolomite, dolomite, graphite, diopside, quartz-diopside, and forsterite marbles identified by their mineral composition. The entire marble sequence has been cut by sills, dikes and veins of gabbro-diabase and pegmatites of a Precambrian magmatic cycle.

Post-magmatic alterations have been widely developed in the Pereval marbles. They are most intensive at the contact of pegmatites with dolomitic and forsteritic marble, where spinel-bearing zones, which originated in a bimetasomatic replacement of pegmatite and marble, occur along with other formations. Spinel occurs in these zones in small light-blue grains and crystals.

Ruby spinel occurs under quite different conditions. In the deposit area, it has been found in forsterite and dolomite-calcite marble, unrelated to pegmatite bodies.

Macroscopically, forsterite marble is a white medium-grained rock, of about 20% forsterite, 40% dolomite and 40% calcite. There is some diopside, with or without the forsterite, and rarely phlogopite and apatite. The rock texture is most commonly massive, with banded varieties. In banded marble, forsterite is concentrated in intercalations, 5 to 20 cm thick, 20 to 40 cm apart. Ruby spinel was found in some of these bands, unevenly scattered throughout the rock. In forsterite-rich intercalations, with 50% forsterite, 20% dolomite and 20% calcite, the spinel content reached 15 or 20% of the volume of the intercalations. A photomicrograph of spinel-forsterite marble is presented in Figure 1-a.

Macroscopically, spinel-bearing dolomite-calcite marble is a white, coarse-grained rock. Coarse calcite crystals enclose fine, rounded and evenly distributed dolomite inclusions, on the average of 30% of the calcite bulk; also crystals and amorphous, frequently replaced, grains of ruby spinel, up to 5% of the carbonate body.

In both types of marble, spinel crystals exhibit a regular octahedral habit, although there are some with irregular faces. The crystals are usually clear, with lustrous faces, almost devoid of fractures and roughness. In the rare instances where there are some fractures, they are evenly distributed over the face, penetrating more or less deeply into the crystal body. The fracture rims are rounded, as if fused, and the fractures themselves are filled with carbonate, forsterite, or phlogopite.

Besides individual crystals, there are crystal growths, step-like octahedrons, and twins according to the spinel twinning law. The size of crystals ranges widely, from fractions of a millimeter to 1.5 cm, with 0.3 to 0.5 cm crystals predominating. A single step-like octahedron has been found to

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[1] Krasnaya shpinel' v mramorakh mestorozhdeniya pereval i yeye vtorichnyye izmeneniya.

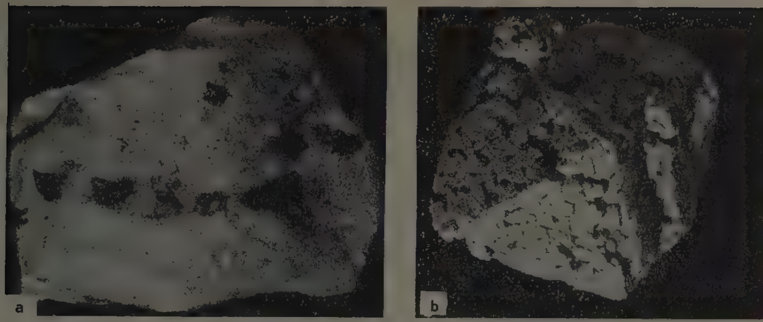

FIGURE 1-a -- forsterite marble (light-colored) with crystals of ruby spinel (dark), natural size; 1-b -- step-like octahedron of ruby spinel; 2.5X.

measure 2 cm, weighing 11 grams (see Fig. 1-b). In color (beautiful red of various hues) and transparency, the Pereval spinel is similar to precious varieties. In addition, rosy, dark red, brownish red, purplish red, and other varieties have been found.

It was believed up to recently that the ruby color of spinel was due to the presence of chromium. According to L. M. Lebedev and N.G. Sumin [8], it is due also to vanadium which they have found to be constantly present in red spinel. In 1955, E. Thilo and R. Sauer [10] studied the $MgO \cdot Al_2O_3$ - $MgO \cdot Cr_2O_3$ system and concluded that it formed an isomorphous spinel series whose color depended on the amount of $MgO \cdot Cr_2O_3$. Thus, the spinel color is always ruby, with the $MgO \cdot Cr_2O_3$ as much as 15%; above that, it changes to green, by way of gray. Thilo and Sauer believe that at the 15% content of that component, one seventh of Al ions has been replaced by Cr ions, so that individual Cr^{+3} ions come in contact only with Al^{+3} ions; above this figure, they contact each other, as well. This view is in accordance with data of K. Fayans [9] who regarded the coloring of a crystalline substance not with relation to specific properties of individual cations but as a result of a definite combination of these cations with anion groups.

It should be kept in mind, however, that the coloring partly due to chromium may be camouflaged by trivalent iron, present sometimes as an isomorphous impurity in spinelids. Thus, according to P.V. Kalinin [5], spinel from the Slyudyanka cliff, containing traces of chromium and 2.18% Fe_2O_3, is light blue, but spinel from the "White Quarry," with 0.82% Cr_2O_3 and lacking the trivalent iron, is red. As analysed by M.G. Zamuruyeva, the red Pereval spinel, from an exposure 72 m from the upper rim of the highway cut turn, contained Al_2O_3, 69.95%; FeO, 1.31%; CaO, traces; MgO, 27.51%; Cr_2O_3, 0.69%; V_2O_5, traces; loss in heating, 0.76%. SiO_2, TiO_2, Fe_2O_3 and $Na_2O + K_2O$ were absent. Bivalent iron is present in both blue and red spinel and apparently has little effect on their coloring.

Table 1

Data of semi-quantitative spectrum analyses

Elements	Blue Pereval spinel	Ruby Pereval spinel	Ruby Aldan spinel
Mn	0,0*n*	0,000*n*+	0,00*n*+
Ga	0,0*n*—	0,00*n*+	0,0*n*—
Mo	0,000*n*—	—	—
V	0,000*n*	0,0*n*	0,0*n*+
Cu	0,00*n*—	—	—
Zn	0,*n*—	0,1	0,0*n*+
Ti	0,00*n*+	0,00*n*	0,00*n*
Co	—	—	0,00*n*
Ni	—	—	0,00*n*+
Mg	*n*+	*n*+	*n*+
Si	0,*n*	0,01	0,0*n*—
Al	*n*+	*n*+	*n*+
Fe	*n*+	0,00*n*+	*n*—
Cr	0,001	0,*n*—	0,0*n*
Ca	0,00*n*	0,0*n*—	0,*n*

Symbols: *n*(—) = 1—3 %; *n* = 4—6 %, *n*(+) = 7—9 %.

Note: Comma represents decimal point.

With ultraviolet rays, $\lambda = 360\ \mu\mu$, the red Pereval spinel luminesces bright-red, but the blue spinel from the same locality and the red Aldan[1] spinel do not luminesce. This

[1] A specimen of white marble, rich in clinohumite and red spinel was kindly put at our disposal by N.N. Pertsev.

lack of luminescence in the last two types may be explained by their high iron content (see Table 1) which is known to be a powerful luminescence damper [1, 3].

Under the microscope, the red Pereval spinel is colorless and isotropic. Its refractive index for sodium light is 1.716 ± 0.002. For mineral impurities, spinel crystals contain dolomite, forsterite, and phlogopite; the most common, however, is corundum which is present either in large hexagonal crystals (as large as 2 mm) or in very fine amorphous grains, evenly distributed in an isotropic matrix or else concentrated in individual microfractures (Fig. 2). The first type of corundum appears to have been contemporaneous with the spinel. The second type -- as will be shown below -- was formed by its replacement.

The Pereval spinel-bearing marbles have undergone a number of alterations. We have identified four types of spinel replacement: diopsidization, phlogopitization, carbonatization, and chloritization.

Diopsidization appears to be the highest-temperature process of spinel replacement. This is a rather uncommon process, being associated exclusively with the outer contact of granite-pegmatite veins. The specimen under study, exhibiting such a phenomenon, contained approximately 50% diopside, 40% calcite, 10% spinel and traces of dolomite. At the contact with calcite, spinel was replaced by diopside which was also developed in fine fractures within the spinel crystals. We believe this diopsidization process to be a result of the addition of silicon liberated in the metasomatic desilication of pegmatite.

The replacement of spinel and forsterite by phlogopite occurred under quite different conditions and apparently at lower temperatures. This process took place in a narrow zone of spinel-forsterite marble immediately contacting the obliquely-cutting veins of pegmatite. The intensity of the process decreases away from the contact and dies down completely 0.5 mm from it. The phlogopitization of spinel and forsterite begins at their contact, with the formation of comparatively coarse phlogopite tablets. Deeper in both minerals, aggregates of very fine phlogopite scales, exhibiting many extinctions, have been formed. Comparatively coarse phlogopite tablets have also been formed at the contact between dolomite and minerals in the process of replacement. Where the contacting spinel and forsterite have been fully replaced, an elongated aggregate of phlogopite scales is seen under the microscope. It is surrounded by coarser phlogopite tablets and is transversely cut by them, thus suggesting the original contact of these two minerals. A photomicrograph in Figure 2 shows a large grain of ruby spinel containing corundum crystals and finer inclusions of that mineral. The spinel is enveloped in the fine-scaled aggregate and surrounded with larger phlogopite tablets.

The phlogopitization of spinel and forsterite is accompanied by an abundant precipitation of

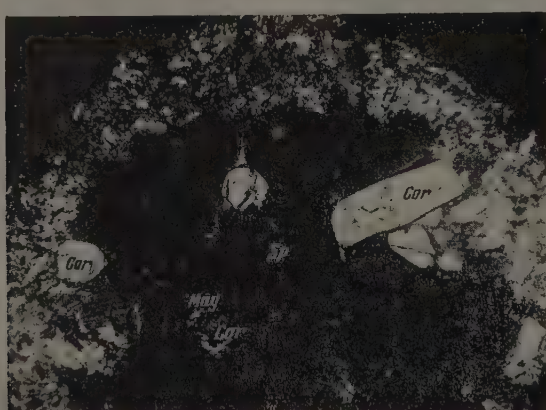

FIGURE 2. Isotropic mass of spinel with corundum crystals (light) and microfractures (dark) filled with secondary corundum and amorphous magnesite.

Sp -- spinel; Cor -- Corundum; Fl -- Phlogopite; Mag -- magnesite. Crossed Nicols, 8X.

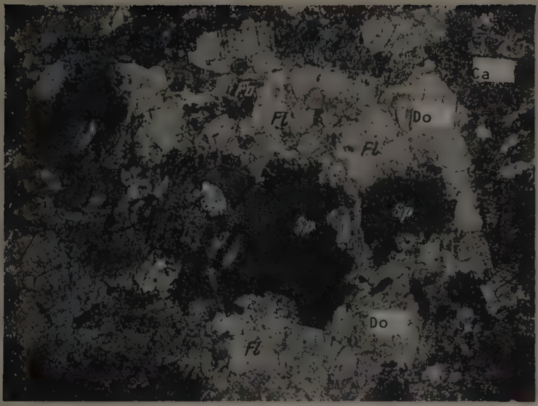

FIGURE 3. Replacement of spinel, forsterite, and calcite by phlogopite and dolomite.

Fo -- forsterite; Do -- dolomite; Ca -- calcite; Sp -- spinel; Fl -- phlogopite. Crossed Nicols, 8X.

magnesium oxide, as witness the dolomite fringes around the minerals in replacement. Photomicrograph Figure 3 shows two spinel grains, replaced with phlogopite on two sides, locally with forsterite relicts, the whole enveloped in a dolomite fringe.

The chemistry of the phlogopitization process is very complex. The formation of phlogopite from spinel and forsterite, in the absence of K-minerals in the rock, suggests that potassium has been brought in by post-magmatic solutions. The replacement of calcite with dolomite was evidently accompanied by the addition of CO_2. Alumina and silica were borrowed from spinel and forsterite, respectively. Both alumina and silica could have fully entered the phlogopite lattice; but magnesium, abundant in both reactive minerals, was not fully expended in the formation of phlogopite and its excess was responsible for phlogopitization of calcite. Iron, present in a small amount in forsterite (2% Fe_2SiO_4, according to the refractive indices), has fully entered the phlogopite. According to a study by P. V. Kalinin [4], the optical constants of phlogopite ($\gamma = \beta = 1.577 \pm 0.002$; $\alpha = 1.541 \pm 0.003$; γ-$\alpha = 0.036$) are typical of a low-iron phlogopite, widely developed in the Slyudyanka marble.

The phlogopitization process in spinel-carrying dolomite-calcite marbles begins at the crystal periphery and proceeds inward, until all of the spinel has disappeared. As a result, the spinel is replaced with corundum while the liberated magnesium oxide is not leached but is used in the dolomitization of adjacent calcite grains. There are instances of a few grains of secondary corundum, intergrown with corundum, developed on the spinel periphery or in microfractures. No corundum has been observed beyond the former spinel crystals. Figure 4 illustrates a large spinel crystal with its periphery replaced with corundum and dolomite. The sharp and straight traces of the spinel faces are plainly visible, as are its replaced edges and a dolomite fringe, developed on calcite about the spinel in replacement. Figure 5 shows a rectangular aggregate of corundum, with a dolomite fringe followed by calcite. The spinel grain appears to have been replaced with corundum, while the liberated magnesium formed dolomite.

The carbonatization process in forsterite-spinel marble runs a different course. Here, in the absence of a direct spinel-calcite contact, the spinel is replaced by magnesite and corundum. The easiest replaced are spinel kernels in phlogopite envelopes, with amorphous spinel relicts fringed successively with fine-grained corundum and amorphous magnesite, within an earlier fringe of phlogopite. In addition, magnesite and corundum form fine micro-veins in fractures of the spinel relicts (see Fig. 2). These micro-veins commonly exhibit a zonal structure, with magnesite in the central part and corundum at the periphery.

Both the magnesitization of spinel and the dolomitization of calcite about the spinel in replacement apparently proceeded under similar thermodynamic conditions. The course of

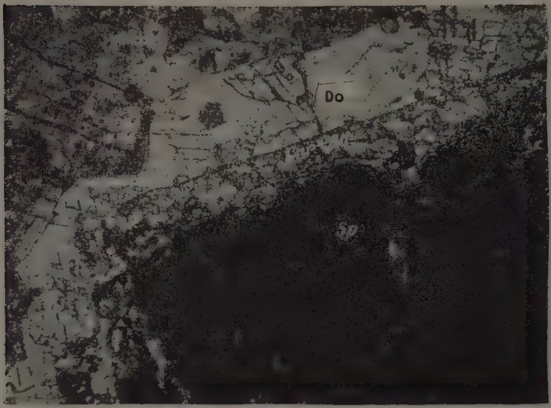

FIGURE 4. Replacement of spinel and calcite by corundum (symbols same as in Figure 3). Crossed Nicols, 16X.

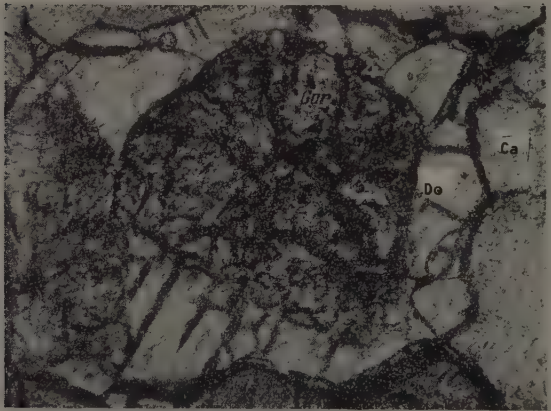

FIGURE 5. Corundum aggregate (Cor) with a dolomite fringe (symbols same as in Figure 3). Crossed Nicols, 32X.

both was determined by the magnesium concentration in the medium and by the circulation intensity of hydrothermal solutions. Judging from the structural relationship between minerals participating in the carbonatization of spinel, the chemistry of this process is rather simple, being a decomposition of spinel into MgO and Al_2O_3. In this process, the silica is inordinately inert, while MgO is either sluggish or inert, depending on the magnesium concentration of the medium. In the first alternative, MgO reacts with calcite and with CO_2 of the solution to form dolomite; in the second, MgO reacts with CO_2 alone to form magnesite.

The phenomenon of the spinel and calcite replacement by corundum and dolomite was first discovered by D. S. Korzhinskiy [7] in metasomatic spinel veins, in the upper Uluntuy rapids (Slyudyanka district). Korzhinskiy believes that corundum has been formed under changing equilibrium conditions, at a lower temperature, along with the carbonatization of spinel, as follows: $MgO \cdot Al_2O_3 + CaCO_3 + CO_2 = CaMg(CO_3)_2 + Al_2O_3$. P. V. Kalinin holds a completely different view [5], in regarding the presence of corundum in

metasomatic veins as a result of the deepest desilication of pegmatite veins.

The spinel chloritization process occurs exclusively in forsterite-spinel marble, where it is superimposed on an earlier phlogopitization process or else is developed simultaneously on spinel and forsterite, where these minerals directly contact each other or are close together. As in phlogopitization, the chloritization of spinel and forsterite is accompanied by an abundant liberation of magnesium oxide, which brings about an intensive dolomitization of the adjacent calcite grains.

Chlorite, judging from its optical properties, ($\gamma = 1.589 \pm 0.002$; $\beta = \alpha = 1.578 \pm 0.002$; $\gamma - \alpha = 0.011$; $+2V = 0$), should be assigned to amesite which contains 58% of the amesite molecule proper ($H_4Mg_2Al_2SiO_9$) and 42% of the antigorite molecule ($H_4Mg_3Si_2O_9$).

The chemistry of chloritization is essentially a hydration of spinel and forsterite with a simultaneous dolomitization of calcite. Principal components necessary for the formation of amesite are found in the reacting minerals. Iron, present in forsterite to the amount of 2% Fe_2SiO_4, apparently is not leached, but enters the amesite molecule instead, where it isomorphously replaces magnesium; while chromium of spinel, 0.69% by weight, isomorphously replaces aluminum.

The chloritization of spinel completes the deep-seated metamorphic cycle for this mineral. Under the subsequent hypergenetic conditions, spinel displays exceptional stability, as witness the well-preserved crystals in a weathered surface of spinel-bearing marble.

Associations of forsterite-spinel-calcite-dolomite and of spinel-calcite-dolomite, formed in a regional-contact metamorphic process, are subsequently acted upon by postmagmatic solutions circulating throughout the weakened zones, and change to other associations -- different stages in the metamorphic evolution of spinel-bearing marble. Typical of these stages is the selective participation of the rock minerals, with the progress of associations presented as follows:

1. Spinel-calcite → diopside.

2. Spinel-forsterite-calcite → phlogopite-dolomite.

3. Spinel-calcite → dolomite-corundum.

3a. Spinel → magnesite-corundum.

4. Spinel-forsterite-calcite → chlorite-dolomite.

Each transition step reflects a change in the thermodynamic parameters. It may be stated that, on the whole, temperature and pressure tend to decrease from the first to the fourth stage. Change in the paragenetic associations, at the first stage, take place at comparatively high temperatures and pressures and are not accompanied by absorption of H_2O and CO_2. Accordingly, this stage may not be regarded as retrograde metamorphism. The other stages are marked by lower temperatures and pressures, with the change in the paragenetic association accompanied by absorption of H_2O and CO_2 (negative thermal effect reactions). Consequently, they are typical of retrogressive metamorphism. The second and fourth stages are characterized by a high activity of water and carbon dioxide; the two third ones, by a high activity of water but a low carbon dioxide activity. The activity of water affects the character of mineral transformations. At the second and fourth stages, they are reactive, which is not true for the two third-stage transformations. One common feature of mineral transformations in a retrogressive metamorphism of spinel-bearing marble, however, is their hysterogenetic character, inasmuch as they all are genetically related to Archaean pegmatites. Moreover, these transformations are realized essentially at the expense of the original rocks, except for the second stage where the transformation runs with an addition of potassium.

We note, in conclusion, that a detailed study of retrograde metamorphism in all varieties of the south Cis-Baikal marbles may lead to valuable conclusions on the conditions and character of this interesting but little-known type of metamorphism.

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Received May 27, 1958

BRIEF COMMUNICATIONS

SOME DATA ON THE EFFECT OF GEOLOGIC CONDITIONS ON THE FORMATION OF THE TERRESTRIAL NEUTRON FLUX[1]

by

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A substantial component of the atmospheric neutron flux are the secondary neutrons of cosmic rays. The flux reaches its maximum in the stratosphere, decreasing at the earth's surface to a low value (20 to 200 neutrons $cm^{-2}day^{-1}$), [?] with slow neutrons predominating. The cadmium ratio, i.e., the ratio of the full neutron flux to that passed through a cadmium layer -- an absorber of slow neutrons -- at the earth's surface, is large, being in the order of 40.

The presence of a neutron flux in the earth's crust has been determined indirectly, from numerous reactions of synthetic transformations (the accumulation of He^3 in spodumene, in the fission of lithium; synthetic fission of AcU with the formation of krypton and xenon isotopes; the accumulation of xenon in ancient telluriferous minerals, and of argon and neon in uraninite; the formation of Pu^{239} and Np^{237} and other neptunium series isotopes in radioactive minerals). Even in the presence of mineral radio-elements capable of spontaneous fission with an emission of neutrons, the formation of the nuclear transformation products cannot be explained by this factor alone.

Direct measurement of the neutron flux in the earth's crust are few. In 1954, Eigster noted that the neutron flux in one of the Swiss tunnels increased approximately by the factor of 30, as compared with that at the surface. Neutrons were registered with plates thickly covered with an emulsion saturated with lithium and boron. The number or events at depth was 4.71 cm^{-2} day^{-1}, which according to Eigster, corresponds to a neutron flux of 608.5 neutrons cm^{-2} day^{-1}. The cadmium ratio was low (2.42), i.e., fast neutrons were present to a considerable extent. In 1954, V.V. Cherdyntsev and V.I. Meshkov discovered sizable variations in the neutron flux [1]. The neutrons were registered with thin-layered plates coated with lithium carbonate or boracic acid. The sensitivity of this method was 2 to 3 times lower than Eigster's. On a glacier, the neutron flux dropped to zero, within the margin of error, rising up to 10 alpha-particles cm^{-1}, which corresponds to about $3 \cdot 10^3$ neutrons cm^{-2} day^{-1}. The cadmium ratio, too, was low (of the order of 2.75).

Inasmuch as the accepted main component of the terrestrial neutron flux is neutrons of reactions (α, n), with beryllium the best neutron-generating substance, we began with the objective of studying the neutron flux in the zones of beryllium mineralization. In addition, we performed a comparative study of the neutron flux in the laboratory (Alma-Ata, 800 m) and in one of the neighboring glaciers (about 3,400 m).

In the neutron registration, we used thick-coated plates whose emulsion contacted (for alpha radiation) a saturated layer of boron carbide (77% of boron). Prior to the study, the plates were processed with an oxidizer (potassium ferricyanide solution), to eliminate the latent alpha-ray images. As in our former experiments [1], the plates were exposed in standard paraffin envelopes, 1 cm thick, which increased somewhat the relative number of slow neutrons. The envelopes were used as protection from moisture which caused a regression of alpha particles. It has been estimated that a single trace from the boron carbide layer to the emulsion corresponds to approximately 100 neutrons passing through the plate. Unfortunately, the sensitivity of plates was uneven, while the accumulation of background particles was too great (about 3 to 10 alpha particles cm^{-2} day^{-1}). Accordingly, the precision of the experiment was not great. The intensity of gamma radiation

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[1] Nekotoryye dannyye o vliyanii geologicheskikh usloviy na formirovaniye neytronnogo potaka zemli.

was measured with a field counter, at all points.

The working procedure in the area of pegmatite veins was as follows. Three plates in individual standard paraffin envelopes were exposed at each point. One of them was placed in a cadmium box with the walls sufficiently thick to absorb all slow neutrons. The envelopes were carefully sealed with paraffin, against moisture. Envelope c (plate of cadmium protection) was placed flat on envelope a, with envelope b alongside. Under such arrangement, difference b — a of traces was determined by the absorption of neutrons coming from above and passing through the cadmium box. The number of alpha particles was measured in a 2 square centimeter area.

A comparison of data for different localities shows a sufficient consistency in the alpha particles' distribution over the plate surface. Control plates, exposed in a field laboratory, gave the rate of the alpha-particle accumulation, 11 cm^{-2} day^{-1}, in the absence of the boron carbide contact (the "background" value) and 7.2 cm^{-2} day^{-1}, with it. The difference apparently lies within the sensitivity range of plates, and the neutron flux could not be determined in the field laboratory. The measurement results are listed in Table 1.

These data reveal an appreciable decrease in the number of traces on plates screened off with cadmium. The average value of b — c is 6.2 alpha particles cm^{-2} day^{-1},

Table 1

Study of neutron flux in the area of pegmatite veins

Location	Gamma activity	Number of alpha traces on plate, for cm^2/day			
		Total	Minus the background	a—b	b—c
Main vein; surface					
Experiment b	1.0	13.3	2.3		
Depth 17 m					
Experiment b	1.2	17.7	6..7	3.6	4.7
Experiment a	—	14.1	3.1		
Experiment c, cadmium protection	—	13.0	2.0		
Greisen vein, rich in beryllium					
Experiment b	0.8	20.2	9.0	4.4	9.7
Experiment a	—	15.6	4.6		
Experiment c, cadmium protection	—	10.5	-0.5		
Watershed vein, rich in beryllium					
Experiment b	0.6	18.8	7.8	3.3	8.3
Experiment a	—	15.5	4.5		
Experiment c, cadmium protection	—	10.5	-0.5		
Placer					
Experiment b	0.75	22.5	11.5	8.3	9.2
Experiment a	—	14.2	3.2		
Experiment c, cadmium protection	—	13.3	2.3		
Point outside the mineralized zone					
Experiment b	1.0	14.0	3.0	-4.0	0.3
Experiment a	—	18.0	7.0		
Experiment c, cadmium protection	—	13.7	2.7		

vhence the slow neutron flux is about cm^{-2} n the number of traces has been observed or plates covered by the cadmium box. The werage value of b − a is 3.3 alpha particles :m^{-2} day^{-1}. It follows, then, that most neutrons hit the plate from above, from the tmosphere, and are not connected with the errestrial flux.

The point is that the "main vein," seems o exhibit a more intensive neutron flux, with lepth.

The study in the area of hydrothermal nineralization was conducted along the same ines, except that plate b was placed normal o plates c and a. Control plates in the field aboratory have shown 3.9 cm^{-2} day^{-1} alpha particles, without the contact with boron preparation; and 2.4 with it. The higher value has been conditionally accepted as that of the "background." As in the previous experiment, no increase in the number of traces has been observed for plates in contact with boron carbide, which means that the neutron flux, in the field laboratory, is beyond measurement. The results are given in Table 2.

Plates, exposed in workings of the southeastern mineralized zone, where the rock activity is low, show only an insignificant neutron flux, impossible to measure. The surface neutron flux was found to be of the same intensity as in the pegmatite vein area.

An abrupt increase in neutron flux has been noted for deep points of the central zone, where it was as high as 171 alpha-traces cm^{-2} day^{-1}, or approximately $1.5 \cdot 10^4$ neutrons cm^{-2} day^{-1}. The number of alpha particles in experiments a and b, at a point in the southern part (with plates normal to each other), differ strongly. It is possible that this indicates an oblique direction of the flux. The cadmium-protected plate, too, exhibits an increased number of neutrons (the cadmium ratio does not exceed 3.4), which suggests a powerful component of fast neutrons.

Two plates were exposed on the glacier; and two more, for control, in the permanent laboratory. All plates were in standard paraffin envelopes, with one at each station with a cadmium protection. The results are given

Table 2

Study of neutron flux in the area of hydrothermal mineralization

Location	Gamma	Number of alpha traces on plate, for cm^2/day	
		Total	Minus the background
Southeastern zone			
Surface, Experiment b	1.2 ± 0.2	11.8	7.9
Depth 60 m			
Experiment a	1.2 ± 0.2	3.5	-0.4
Experiment c, cadmium protection	−	4.5	+0.6
Depth 110 m			
Experiment b	0.8 ± 0.2	7.2	3.3
Experiment a	=	3.4	-0.5
Experiment c, cadmium protection	−	5.3	1.4
Central zone			
Surface			
Experiment b	1.4 ± 0.2	6.6	2.7
Experiment a	=	9.8	5.9
Experiment c, cadmium protection	−	3.4	-0.5
Depth 60 m; Southern part			
Experiment b	5.0 ± 0.4	175	171
Experiment a	=	38.4	34.5
Experiment c, cadmium protection	−	54.4	50.5
Same; Northern part			
Experiment b	3.1 ± 0.3	40.8	36.9

Table 3

Measurement of neutron flux in a stationary laboratory and in a glacier

Locality	Number of alpha traces on plate, for cm^2/day
Laboratory, 800 m	
Plate without cadmium protection	7.5 ± 0.5
Plate with cadmium protection	7.5 ± 0.5
Glacier, 3,400 m	
Plate without cadmium protection	7.5 ± 0.5
Plate with cadmium protection	7.5 ± 0.5

in Table 3.

No slow neutrons have been observed at either location (fewer than 50 neutrons cm^{-2} day^{-1}). Nor has any intensification of the fast neutron flux been found at the glacier's elevation. This means that any change was beyond the resolution power of our method.

SUMMARY

No intensification of the neutron flux has been detected in the area of the pegmatite beryllium mineralization. Slow neutrons predominate in the flux, and they come chiefly from the atmosphere, as is apparent from experiments with cadmium protected plates. Fast neutrons are beyond measurement (by our method).

No measurable flux of slow neutrons has been detected, either in the glacier or in the permanent laboratory. A sizable increase in the neutron flux has been detected for a beryllium mineralization area with a stepped-up activity. This probably was due to two factors: a sufficiently strong flux of alpha particles and the presence of a neutron-generating substance.

This locality is the only known one where an increase in the terrestrial component of neutron flux lends itself to an interpretation by geologic conditions.

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Received December 19, 1956

THE COMPOSITION AND STRATIGRAPHIC SIGNIFICANCE OF THE UPPER DEVONIAN PELECYPOD ASSEMBLAGES IN THE CENTRAL VOLGA-URAL PROVINCE[1]

by

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Pelecypods make up one of the most common fossil groups in the Upper Devonian of

[1] O sistematicheskom sostave i stratigraficheskom znachenii verkhnedevonskikh peletsipod tsentral'noy chasti Volgo-Ural'skoy oblasti.

:he central Volga-Ural province. Despite that, :he stratigraphic significance of this faunal group still is not well known. This is chiefly because most of the fossil shells are poorly preserved and commonly difficult to identify generically and specifically. Generally, they are dwarfed forms, occasionally (in shale) strongly attenuated, flat, as if compressed (e.g., Pterochaenia, Posidonomya, Nuculana).

As a result, many students doubt the feasibility of using this faunal group as index fossils.

A study of pelecypods from this area has shown their definite stratigraphic value. It has been determined, for instance, that besides forms which have a wide vertical range, there are a number of species associated with smaller stratigraphic subdivisions. A study has been made of the most important assemblages characteristic of a given geologic interval. They comprise the bulk of Upper Devonian pelecypods, best developed in the middle Frasnian substage.

Systematically, this assemblage is represented by the following genera: Buchiola, Pterochaenia, Paracardium, Ontaria, Nuculana, Posidonomya, Praecardium (?), Lunilucardium, Nucula, and Conocardium.[1] Along with linguids, ostracods and pteropods, these genera form an index assemblage for the widely developed "Domanick" type rock.

Among other pelecypod groups of the central part of the Volga-Ural province, there are representatives of the following genera: Avicula, Aviculopecten, Modiella, Schizodus, Paracyclas, Paraptyx, and occasionally Mytilarca, Modiomorpha (?), Pecten and others. However, the above-named genera are less numerous and less valuable as index fossils with the exception of certain representatives of genus Aviculopecten). For this reason, they have not been subject to special study. Moreover, many of them were described before, in works of B.V. Nalivkin [3] and A. Zamyatin [2], from the western slopes of the Southern Urals and Timan. There is every reason to believe that they will become more important, with more detailed study.

In evaluating the volume and boundaries of genera we considered the morphology of the shells, their stratigraphic and facies position, and their geographic distribution.

As a result of processing our own material from boreholes of the Tatarya, Bashkirya and adjoining provinces, and of an analysis of published data, we have identified 25 species of 10 genera and five families. Some of these forms, in association with other species, are typical of definite stratigraphic horizons, only. The most important are representatives of genus Buchiola, which account for some 14 forms. Because of their wide distribution, the frequency of occurrence, and fair preservation, they are foremost among the index pelecypod fossils.

The study of Upper Devonian pelecypods began with the Frasnian substage where representatives of this group were first noted in Kynovsk deposits. The fauna from this horizon has been much less studied than that from the upper horizons, because of the scarcity of well cores and of the shells themselves, and their poor state of preservation. Identified here were fairly numerous representatives of Pterochaenia fragilis Hall, a few buchiolids different from younger forms but not identifiable because of their poor preservation, and isolated Pterochaenia trigonalis Prok. nom. msc. -- a form best developed in Sargayevsk limestone. In addition, there are forms belonging to genera Avicula, Aviculopecten, and Schizodus (?), usually absent higher up in the section.

Is should be noted that the presence of genera Avicula and Aviculopecten obviously points to a connection of this basin with the Uralian geosynclinal province where these forms are widely developed.

The pelecypod assemblage was vigorously rejuvenated in the Sargayevsk formations of gray argillaceous limestone, mudstone, and dark calcareous shale and limestone of the "Domanick" type. At the same time, the quantitative ratio of species within genera was increased sharply. Present here are Pterochaenia trigonalis Prok., nom. msc., Paracardium doris Hall, Buchiola platicostata Prok., nom. msc., Nuculana tenuilineata Prok., nom. msc., etc. This assemblage has been definitely traced in many boreholes, without ever continuing into an upper horizon. In addition, there are forms with a wider vertical range: Pterochaenia fragilis Hall and Ontaria articulata Münst., with occasional Lunulicardium and Praecardium (?).

Thus the Sargayevsk pelecypod assemblage has an aspect of its own, pinpointing its position in the section.

The "Domanick" pelecypod assemblage is the richest and most diversified, containing over a half of all forms known from the Upper Devonian of the subject area, with representatives of genus Buchiola predominating. Most typical are Buchiola halli Clarke, B. scabrosa Clarke, B. prumiensis Stein., B. eifeliensis Beush, etc. Along with specifically Domanick forms, there are a number of

[1]The genera are listed in the order of decreasing abundance.

others, present in Mendym deposits, such as Buchiola retrostriata Buch., B. lupina Clarke, B. angolensis Clarke, B. palmata Goldf., B. snjatkowi Zam., B. obliqua Prok. nom. msc., etc. At the same time, the number of forms persisting from lower horizons is comparatively small, among them Pterochaenia fragilis Hall, Ontaria articulata Münst., Lunulicardium sp, Praecardium (?). Thus the Domanick-Sargayevsk boundary, based on the pelecypod fauna, is fairly sharp. This is less true for the Domanick-Mendym beds.

The Mendym pelecypod assemblage, in general, preserves its Domanick aspect, except that it is considerably poorer. Buchiola ishacaja B. Nal., B. bellula Prok., nom. msc., Conocardium appear here. B. snjatkowi Zam., a form appearing at the close of Domanick time, is rather poorly developed. No pelecypods have been observed in the Askynsk formation.

Thus, three groups of species emerge in the Frasnian pelecypod assemblage: 1) typical of Sargayevsk deposits; 2) typical of the Domanick-Mendym sequence, but best developed in the Domanick; and 3) species with a wide vertical range.

Pelecypods are very rare in Famennian deposits, the most common being Posidonomya venusta Münst and P. eifeliensis (Frech.).

Considering that these forms are widely distributed, geographically, although not observed anywhere in Frasnian deposits, their stratigraphic significance in the differentiation of Frasnian and Famennian carbonate rocks is obvious.

All of the above is a convincing proof that doubts of some investigators as to the stratigraphic value of this faunal group are not justified. It may be stated unequivocally that the degree of precision in the differentiation of the Upper Devonian of the Second Baku, by means of pelecypods, is in places not inferior to that based on brachiopods and other proven groups.

A study of the systematic composition and vertical distribution of pelecypods has shown that a number of middle Famennian forms are common also to the correlative Naples beds of North America.

This corroborates the view of many students on the identity of the Domanick facies fauna from Southern Timan [2, 7] and western slope of Southern Urals [3] with the Naples fauna. Such a wide geographic distribution of Pterochaenia fragilis, Buchiola retrostrata, B. lupina, B. halli, B. palmata, etc. -- from the Urals western slope to North America -- suggests a connection between these two basins and is not in contradiction with B.V. Nalivkin's view [4] of a pseudoplanktonic habit of these animals. However, the occurrence of pelecypods together with pteropods, goniatites, and orthoceratites is more indicative, to our way of thinking, of a planktonic mode of life for this faunal group, as was hinted before, by D.V. Nalivkin [5] and S.S. Ellern and Ye. Ye. Ivanov [6].

It appears that Upper Devonian pelecypods of the central Volga-Ural province were a peculiar group of marine animals adapted to the special Domanick sedimentary conditions of shallow, poorly aerated marine basins.

With their comparatively uniform generic composition, pelecypods of these facies were subject to considerable changes in time, as expressed in specific assemblages.

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Received December 31, 1957

REVIEWS AND DISCUSSIONS

TUFFACEOUS LAVAS AND IGNIMBRITES (ON THE OCCASION OF THE "TUFFACEOUS LAVAS" PUBLICATION)[1, 2]

by

Ye. F. Maleyev

The objective of this publication of the Volcanology Laboratory, Academy of Sciences, U. S. S. R., is very timely and complex: the problem of tuffaceous lavas and ignimbrites. Therefore, such should have been its title.

The collection contains papers by such expert volcanologists as V. I. Vlodavets, V. P. Petrov, M. A. Favorskaya, I. M. Volovikova, and B. L. Rybalov. Some of the authors deny the possibility of the formation of ignimbrites, as understood by Fenner and Marshall, while the others cite evidence of such formations. V. N. Petrov recognizes the existence in nature of both tuffaceous lavas and ignimbrites.

It may be concluded on the basis of the evidence cited that ignimbrites do exist in nature, along with tuffaceous lavas. However, the authors broaden the meaning of the term "tuffaceous lava" to include not only lava flows with pyroclastic material but also rocks consisting of extrusive fragments caught in the lava, either before or during the outflow.

At the same time, the authors' interpretation of the term, "ignimbrites," is too narrow.

Ignimbrites are rocks formed of fragments which were erupted from a volcano and then cemented by fusing. Such rocks are supposed to be formed from erupted incandescent clouds, such as produced by Mt. Pelée.

No ignimbrites were formed, however, either in the 1912 Mt. Pelée eruption or in the numerous eruptions of the Kamchatka volcanoes. There is no necessity, therefore, to tie the formation of ignimbrites to incandescent clouds. Tuffaceous lavas and ignimbrites differ from each other, in our opinion, in one very important feature: the nature of cementation. Tuffaceous lavas are volcanic clastics cemented with lava, while ignimbrites are pyroclastic rocks consisting of fused (based together) fragments of lava, glass and crystals. They have no cement. Ignimbrites may include "lada" of Indonesian volcanoes, "piperno" from the vicinity of Naples, "Aso lavas" of the southern Japan, agglutinates, etc. Several types of tuffaceous lavas may be distinguished, depending on the manner of their formation -- whether they are lava flows, volcanic necks, explosive vents, or the apex portions of craters.

There is no need of attaching different names to different kinds of tuffaceous lavas, regardless of whether they make up a neck, dome, flow, dike, stock, etc.

Applying a single term, "tuffaceous lavas," to all volcanics consisting of extrusives cemented with lava is unavoidable in geologic mapping. This necessity arises from the fact that identical lavas may have different origins. The nature of this origin can be determined only after a detailed study of the region, which, in turn, calls for an identification and description of rocks.

Thus, the name, "tuffaceous lavas," may be retained for rocks of different origin but consisting of extrusive fragments cemented with lava.

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CHRONICLE

THIRD SESSION OF THE INTERNATIONAL ASSOCIATION FOR THE STUDY OF THE INTERIOR OF THE EARTH'S CRUST (SEPT. 16-26, 1958) (CENTRAL MASSIF OF FRANCE)[1]

by

Ye. V. Pavlovskiy and L.V. Pustovalov

The International Association for the Study of the Interior of the Earth's Crust, created at the XIX Session of the International Geological Congress, Algiers, has as its objective the study of the most ancient formations, by direct observation. For this purpose, the Association holds sessions in those regions containing such formations.

The preceding Second Session took place in Scotland, in 1957, with representatives of the U.S.S.R. attending for the first time (See Izvestiya of the Academy of Sciences, U.S.S.R., Ser. Geol., no. 1, 1958).

The Third Session, held in the Central Massif of France, September 16-26, 1958, was attended by a delegation from the Academy of Sciences, U.S.S.R., consisting of Academician P.A. Borukayev of the Kaz. S.S.R. Academy of Sciences, Professor-Doctor Ye. V. Pavlovskiy (Geological Institute, Academy of Sciences, U.S.S.R.), and Doctor I. Kh. Khambrayev (Academy of Sciences, Uzb. S.S.R.).

The total attendance was 35 geologists, with local French geologists participating in field trips and special sessions, thus making the total number of participants as high as 60 or 70.

In attendance were representatives of Australia, Belgium (including the Belgian Congo), Holland, Soviet Union, France (including Algeria and Madagascar), and Switzerland. Among them were well known scientists such as Professor P. Micheau (Belgium), M. Rocque and P. Lapadue-Arg (France), W. Newvenkamp (Holland), A. Streckeisen (Switzerland) and others. The organizer, as in the year before, was P. Micheau, General Secretary of the Association.

The purpose of the session was to acquaint the participants with the geology of the Central Massif of France, by means of field trips across the Massif, from west to east. The leaders were Professors M. Rocque and Lapadue-Arg, of the Clermont-Ferrand University and F. Forrestier, M. Chenevoix and J. Peterlongo.

The itinerary had been so planned as to demonstrate the principal elements of the Massif's metamorphics, their tectonics, and the main types of magmatic rocks, differing in composition and age. The trip started in Limoges and proceeded by way of Clermont-Ferrand, Le Puy, and St. Etienne to Lyon, in the Rhone graben which separates the Central Massif from the Alpine zone. The length of the trip was over 2,000 km.

It is well known that the Central Massif of France is a major block, a projection of the Alpine platform basement. It borders on the Paris basin, in the north; on the Aquitanian basin in the south and southwest; and on the Alps, in the east. In most general features, the Massif is a junction of two main Upper Paleozoic fold systems -- the Armorican (northwest) affecting the western part of the Massif, and the Variscan (northeast) characteristic of its eastern part. On the whole, the structure of the Central Massif looks like the letter V, open to the north. However, this simplicity disappears in the details. Specifically, Variscan fold trends of second and higher orders are present in the western half of the Massif, in Limousin Department. It is possible that the change in the trend of folding of ancient

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[1]Tret'ya sessiya mezhdunarodnoy assotsiatsii po izucheniyu glubinnykh zon zemnoy kory (16-26.9. 1958) (tsentral'nyy massiv frantsii).

crystalline schist of the Massif is due to pre-Hercynian "kernels" which are locally exposed within the Hercynian fold zone.

A number of faults crisscross the folded crystalline schist and associated granite massif. The most significant is the nearly straight rift zone, le Grand Sillon Houiller, a coal-bearing zone trending southwest-northeast and cutting the Massif in two nearly equal parts. An abrupt change in the trend of folding takes place here, from the Amorican to the Variscan. Some investigators regard Le Grand Sillon as a zone of "schaarung," a twisting of folds; while some others liken it to the African rifts.

The second major break is represented by the so-called Argentat rift, west of Le Grand Sillon. The Argentat rift is represented by a wide zone of mylonitonized rock, trending northwest over almost 200 km. According to E. Raguen, it represents the front of an ancient pre-upper Carboniferous (pre-Stephanian) overthrust.

Most of the Central Massif is made up of mica schist, gneiss, and migmatite, associated with the granite massif. Granites are very widely developed, taking up a little less than half of the Massif area. It has been long known that the crystalline schist and granite are older than uoper Carboniferous (Stephanian) deposits which are locally represented by lacustrine coal measure. The upper Carboniferous rests with a sharp unconformity upon crystalline schist and granite. Upper Carboniferous deposits, although intensively folded (Saalian phase), have not been metamorphosed. Sedimentary series, older than upper Carboniferous, are developed on the edges of the Massif. For instance, a series with organic remains and assigned to the lower, middle, and partly upper Paleozoic (from Lower Cambrian to Visean) is known from the Black Mountains area, in the south. In the north, Upper Devonian and Visean sedimentary series are present in the district of Morvan and in the Bourbonnais Mountains.

It appears from the work of J. Bergeron in the Black Mountains (1911) and of A. Michel-Levy in Morvan (1913) that Paleozoic pre-Stephanian deposits of these two regions change -- to the north and south, respectively -- to the Central Massif crystalline schist. If this is true, these crystalline schists are nothing but metamorphosed Paleozoic deposits, Cambrian to Visean in age. The sequence was metamorphosed during the subsidence of the geosyncline and then folded in a post-Visean but pre-Stephanian time (Sudetian phase of the Hercinids). Such being the case, there are no traces of ancient -- Precambrian and Caledonian folds -- within the Central Massif. However, the problem of the age of folding of crystalline schist is no longer that simple. The work of the last two decades (J. Jung and M. Rocque) has shown that the stratigraphy of the crystalline schist is more complicated than believed before; the metamorphic process, too, turns out to be very complex. A number of zones of progressive metamorphism have been identified: upper mica schist (main facies of sericitic schist), lower mica schist (with biotite), upper gneiss (bi-micaceous), and lower gneiss (with biotite and sillimanite). All these metamorphic zones have been migmatized to a different degree.

In the north (at Morvan), the migmatization front lies below the lower gneiss; it rises gradually, to the south, to reach the upper mica schist, in the Black Mountains region.

This differential metamorphism of crystalline schists apparently is related to their stratigraphy (J. Jung). In the Black Mountains, crystalline schists are conformably overlain by a continuous sedimentary sequence, from Cambrian to Dinantian. This section corresponds to the Bergeron--Michel-Levy concept of an ancient Hercynian series. It should be noted that this series has also been assigned to the so-called Pyrennean type. Here, the associated migmatites are coarse-grained and their front reaches up to the upper mica schist zone. The picture is quite different in Morvan, to the north. There, non-metamorphosed Devonian and Dinantian deposits rest unconformably upon crystalline schists which, consequently, must be pre-Hercynian (Auvergne type). Migmatites of that series are represented by cordierite anatectites (Obusson type), finer-grained than the Black Mountains migmatite. Anatectites are localized in the lower part of the lower gneiss and are older than the Black Mountain migmatite.

J. Jung believes that the Auvergne-type series is an older, Precambrian formation which forms a massive block in the north of the Central Massif, surrounded by younger, Hercynian folding of the Pyrennean type. It was these Hercynian folds that had produced the classic structure with its Armorican and Variscan branches.

The latest observations make it possible to refine this concept. According to M. Chenevoix, a metamorphic series is present in Limousin, which he assigns to the Auvergne type because it carries Obusson-type anatectites in the lower portion of the lower gneiss. This series, however, is polymetamorphic, since it has been affected by the process of secondary stratified migmatization, more recent than the Auvergne-type migma-

lzation. In the upper course of the Allier, '. Forrestier discovered similar phenomena. le has shown that the front of the secondary tratified migmatizations crosses the metanorphic boundaries within crystalline schist, vhile that of the ancient (primary) migmatiation has the same configuration, in plan, s the progressive metamorphism zones.

J. Peterlongo has proven the existence of wo non-contemporaneous series of crystalline schists in the Lyonnais Mountains. So ar, this is the only place where a clear onnection has been established between the ncient Auvergne block and Hercynian folding. The lower, Lyonnais Mountains series, is uite similar to the Limousin series, in its ouble migmatization and "diaphthoric" pheomena. This lower series is overlain unconormably by the Brevennes series, whose lercynian age has been arrived at by correlation of its extrusives with the Devonian xtrusives of Morvan.

The side trips were concentrated in three reas (Limousin, Upper Allier, and Lyonais) where new and important discoveries ad been made by M. Chenevoix, F. Forestier, and J. Peterlongo, as mentioned bove. A direct study of exposures by the articipants led to appreciation of the preision of observation and of the logic of eduction of these three young students of he Central Massif.

Mush attention was directed to the assorted nagmatic rocks, chiefly granitoids, and heir relationship with the enclosing rocks. Ve had a chance to get acquainted with masifs of anatectitic granite, usually associated vith the very base of a standard crystalline chist section. P. Lapadue-Arg proposed the ame, "basement granite," for these rocks. Their contacts with the enclosing gneiss are ery vague. The granites are marked by heir heterogeneity, obvious even in hand pecimens, and by an alteration of zones of nelanocratic and leucocratic components. The granites carry numerous gneiss incluions, of various form and size. Cordierite s usually present in the granites. Besides arge bodies of anatectic granite, there are omparatively small bodies of the same omposition, which suggest the presence of nany small anataxis hearths. Peculiar spheoidal granites are associated with anatectic granites, in the vicinity of the village of Chataignet. The spheroids consist generally f fibrous oligoclase with a center of either gneiss, granite, or the cementing rock.

The participants were shown rocks closely elated to migmatites -- sub-autochthonous granites in the lower stages of migmatization. natectic granites are called here subutochthonous where they have been squeezed out of their original hearths, in intrusions and even dikes in the emplacing gneiss. The latter-stage granites are more homogeneous and finer-grained than the anatectic granite. They usually form small bodies and veins. We also saw relatively younger intrusive granites and associated representatives of the vein series -- leucocratic granite, microgranite, and lamprophyres (kersantite, spessartite, and minette).

Less attention was paid to basic and ultrabasic rocks (eclogite, peridotite, amphibolite and gabbro) which have but a limited distribution in the Central Massif.

In summarizing the results of the field trips, the high scientific standards of the French geologists should be commended. It must be stated, however, that their emphasis has been on the petrography rather than on the lithological and stratigraphic differentiation of paleontologically barren formations of the crystalline schist series. It is understandable, then, that geologic maps of the Central Massif are actually petrographic maps. As such, they do not adequately reveal the tectonics of ancient metamorphic formations.

Determination of the absolute age of the ancient Central Massif rocks is at an initial stage, for the time being.

One can agree with M. Rocque that one of the important objectives in the study of the Central Massif is the age determination of the Lyonnais series, or, in a broader sense, of the Auvergne sequence. Their upper age limit is undoubtedly pre-Devonian. J. Jung may be right in assigning them to the Precambrian. But the upper part of the Limousin upper ancient series is similar to the Cambrian and Silurin section of the Vendée, which makes probable its lower Paleozoic age.

Another specific, but no less important, problem is the significance of the dual migmatization of the Auvergne-type series, and the age of this process. There is a tendency to connect the first migmatic stage with pre-Devonian metamorphism; and the second -- with the Hercynian. The latest study in the Lyonnais area has shown, however, that repeated metamorphism occurred prior to the deposition of the Brevennes Devonian series, and is therefore pre-Devonian.

A number of meetings were held during the Session. At the first one (September 16, 1958) Academician Polkanov was unanimously elected President; Professors W. Newvenkamp (Holland) and E. Deneyer, Vice Presidents. The constitution of the Association was finally agreed upon, and the Association

Bureau elected. Academician Polkanov read his paper on the geology of isotopes, and L.V. Pustovalov showed a film of the Second Session, held in Scotland, 1957. The results of the trip were discussed at the final meeting. It was decided to hold the 1959, or fourth, session in West Germany (in the Schwartzwald and Odenwald) and in France (in the Vosges).

The participants got acquainted with the activities of Clermont-Ferrand University, its geologic museum and laboratories; they also visited the Puy de Dôme summit, one of the most interesting Quaternary volcanoes. The session was covered by the provincial and Paris press.

In Paris, after the session, the Soviet delegation was kindly shown around the Sorbonne, by French scientists P. Laffitt, N. Menshikov, and M. Orsel. They also visisted the Museum of Natural Sciences and Bureau de Recherches Géologiques, Géophysiques et Minières.

A cordial atmosphere prevailed during the entire stay of the Soviet delegation in France.

The delegation returned to Moscow on October 2, 1958, after 18 days in France.

PERMIAN DEPOSITS IN THE DONETS BASIN[1]

by

V.G. Alekseyev and M.L. Levenshteyn

In his paper, Stratigraphy of Permian Deposits of the Donets Basin (Izvestiya, Ser. Geol., no. 7, 1956), L.P. Nesterenko proposed a new stratigraphic classification of these deposits, on the basis of his own new data.

His very detailed differentiation of the Donbas Permian facilitated the solution of many complex practical problems of prospecting for some useful minerals associated with it (gypsum, ceramic clays, etc.). This is especially true for salt deposits, where the new stratigraphic ideas have been tried on the Artemovsk deposit.

The technical production board of the "Artemuglegeologiya" trust recommended Nesterenko's classification in 1957 to all organizations engaged in exploration and production from the Permian of the Donbas and adjacent areas.

[1]K voprosu o stratografii permskikh otlozheniy Donetskogo basseyna.

At the same time, it was recognized that the usage of old formation names in the new stratigraphic framework was unfortunate.

The old formation names -- the Cupriferous, Limy-Dolomitic, Salt, and Sandy-Conglomeratic -- reflect their lithology which, according to Nesterenko's study, is not typical of them. On the other hand, all Permian formations of the Donbas, except for the Sandy-Conglomeratic, have had their volume changed to various extent. This calls for a new nomenclature.

A commission, created for this purpose, and with the author of the new classification participating, proposed the following changes:

Old	New
Sandy-Conglomeratic, (P_2?)	Dronovsk,* P^{dr}
Salt, P_1^s	Artemovsk, P_1^{art}
Limy-Dolomitic, P_1^c	Nikitovsk, P_1^{nik}
Cupriferous sandstone, P_1^{gr}	Klinovsk, P_1^{kl}

*Offered earlier by L. Lungersgauzen.

The Cupriferous sandstone formation was renamed the Klinovsk because of its wide development and good exposures near Klonovoye village, Artemovsk district. It was here that ancient copper mines were found (Nosov II, 1865).

The Limy-Dolomitic formation was renamed the Nikitovskaya after the type section in that area (near Nikitovka station, the Donets Railroad), where it has been studied in detail, and the associated dolomites have been extensively quarried.

The Salt formation was renamed the Artemovskaya after its type section in the Artemovsk area, Stalinsk oblast', with its well known salt mines.

The Sand-Conglomerate formation has not been adequately studied as yet; accordingly, it has been left almost intact. It should be given one of the present geographic names -- Dekonsk or Dronovsk. The latter is preferable, because it is more familiar. The suggestion of L.P. Nesterenko and G.I. Kireyeva of a lower Permian age for this formation and its conformable position upon the Salt formation has been left moot.

THE FIRST ALL-UNION CONFERENCE ON THE SCIENTIFIC PRINCIPLES OF PROSPECTING FOR BURIED ORES [1]

The first All-Union Conference on Pros-cting for Buried Ores took place in Mos-w, November 18-24, 1958. It was attended representatives of the Academy of Sci-ces, U.S.S.R., system of geologic insti-tions as well as of the Union Academies, e Ministry of Geology and Conservation of ineral Resources, geologic schools of high-learning and of geology departments of iversities, and by field workers of the eological Service, U.S.S.R.

Foreign scientists, representatives of the oples' Democracies, and experts in the eld of ore deposits participated in the Con-rence.

Fifty papers were read on the geology of es, among them papers by Academicians G. Betekhtin and D.S. Korzhinskiy, Cor-sponding Members O.D. Levitskiy, V.I. nirnov, F.V. Chukhrov; Minister of Geol-y P. Ya. Antropov; F.I. Wolfson, Ph.D. eological Mineralogical Sciences; L.I. ikin, Candidate Geological Mineralogical Sciences; E.A. Radkevich, Ph.D. Geological Mineralogical Sciences, and many others. Problems of prospecting for blind ores and of mineral zonation were considered, among other topics.

The papers were followed by a discussion period.

[1] Pervoye vsesoyuznoye soveshchaniye po razrabotke uchnykh osnov poiskov skrytogo orudeneniya.

TO THE READER

The voluminous correspondence of Academician V.A. Obruchev is to be collected for the purpose of a study of various aspects of his scientific and social activities. Persons in possession of his letters are requested to turn them in to the Historical Section of the Geological Institute of the Academy of Sciences of the U.S.S.R. (Moscow, B-17, Pyshevskiy per. 7).

Upon completion of the study, the collected letters will be transferred to the Archives of the Academy of Sciences of the U S.S.R.